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Journal of Sedimentary Petrology

A Publication of the Society of Economic Paleontologists and Mineralogists
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NOTICE

Articles submitted for publication to the *Journal of Sedimentary Petrology* should have a prefatory abstract and sufficient postage should be sent by the author to pay the return of the manuscript as it has been found that many articles require more or less revision before publication.

Persons who are not members of the Society of Economic Paleontologists and Mineralogists or are not subscribers to the *Journal of Sedimentary Petrology* but who desire to publish an article in the *Journal*, should have some member of the Society of Economic Paleontologists and Mineralogists or some member of the American Association of Petroleum Geologists sponsor the article. Members of these Societies are given preference in publication.

JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 19, No. 2, PP. 51–70
FIGS. 1–5, AUGUST, 1949

SORTING OF SEDIMENTS IN THE LIGHT OF FLUID MECHANICS

DOUGLAS L. INMAN

(Contribution from the Scripps Institution of Oceanography, New Series, No. 427. This work represents results of research carried out for the Office of Naval Research of the Navy Department under contract with the University of California.)

ABSTRACT

Observations show that a physical relationship exists between the sorting, skewness and median diameters of sediment samples. Investigation of the fluid mechanic principles involved in the transportation and deposition of sediments indicates that there are three primary controlling factors; the degree of bottom roughness, the settling velocity, and the threshold velocity. Analysis of sedimentary processes in the light of these controlling factors indicates that the relationship of sorting and skewness to the median diameter of a sediment may furnish a clue to the conditions under which the sediment was transported and deposited.

INTRODUCTION

It has frequently been observed that both the sorting and skewness coefficients from a suite of sediment samples are a function of the median diameter of the sample. This fact is well illustrated by Hough's (1942) samples from Cape Cod Bay, by Shukri and Higazy's (1944) Red Sea sediments, and by Krumbein and Aberdeen's (1937) Barataria Bay sediments. In all cases, samples having median diameters near .18 mm are the best sorted; the sorting coefficient increasing in magnitude for median diameters greater or less than .18 mm. The Cape Cod and Barataria Bay samples also show a definite relationship between the skewness coefficient and median diameter. In addition, the dependency of sorting and skewness coefficients upon median diameter has been observed frequently by the author in the analyses of beach and bottom sediments in the southern California region. Visual best fit curves of the plots of sorting coefficients and logarithms of the skewness coefficients versus median diameters for the Cape Cod Bay, Barataria Bay and Red Sea sediments, and for some southern California beach sands, are shown in figure 1.

The description of sediments by the parameters,[1] geometric median diameter Md, coefficient of sorting So, and coefficient of skewness Sk, (or by their equivalent in phi units) is a statistical description analogous to the median, standard deviation, and skewness of mathematical statistics. These parameters are based on the quartile measures Q_1, Md, and Q_3 obtained from the cumulative frequency curve of the mechanical analysis of the sediments. Md is defined as the diameter in millimeters of the middlemost member of the cumulative frequency curve; such that 50% by weight of the sample is larger and 50% smaller. The first quartile measure, Q_1, is that diameter having 75% of the sample smaller and 25% larger; the third quartile, Q_3, is that diameter having 25% of the sample smaller and 75% larger.

The sorting coefficient (Trask, 1932) is defined as

$$So = \sqrt{\frac{Q_1}{Q_3}}$$

and the skewness coefficient as

$$Sk = \frac{Q_1 Q_3}{Md^2}.$$

Both So and Sk are dimensionless, and while physically they appear to be related to the median diameter, this relationship is not implied in their definitions.

[1] All symbols are defined in "Notation and Dimensions" at the end of the paper.

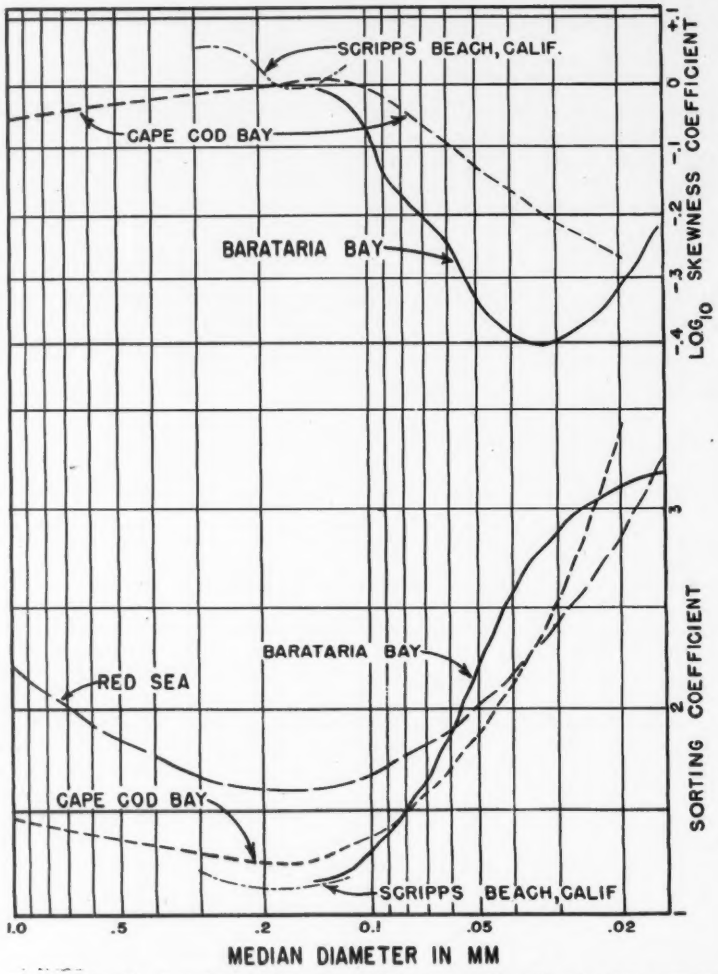

FIG. 1.—Relationship of sorting and skewness coefficients to median diameter of Cape Cod Bay, Barataria Bay, and Red Sea sediments, and Southern California beach sands.

If the histogram of a sediment is skewed towards the fine grade sizes (i.e., the median is finer than the mode) *Sk* will have a value less han 1.0; if it is skewed towards the coarse grades *Sk* will be greater than 1.0. *Sk* will have a value of unity for a symmetrical size distribution of sediments.

The values and interrelationships of these statistical parameters are a complex function of the nature and strength of the transporting agent, the amounts and types of sediment available for transport, the distance transported, and the resulting boundary layer conditions between the transporting medium and the bottom.

It is the purpose of this paper to examine these relationships in the light of fluid mechanics.

ACKNOWLEDGMENTS

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HYDRODYNAMIC RELATIONS

Some of the more pertinent relationships of the boundary layer theory are discussed below. Readers interested in a more complete development are referred to general fluid mechanics texts (for example, Rouse, 1938).

Boundary Layer: Meterologists have investigated boundary layer problems pertaining to the atmosphere and hydraulic engineers have conducted experiments on such problems in flumes, but the results must be modified before being adapted to the ocean or to natural streams.

In general, natural water flows are inherently turbulent in their upper reaches. It should be remembered that while it is customary to describe a turbulent flow by the average velocity $\bar{u}$, the velocity at any instant, u, is the vector sum of the average velocity and a random fluctuation related to turbulence. This may be written

$$u=\bar{u}+u'. \tag{1}$$

The flow conditions near the boundary surface can be considered by examination of a vertical section based at the bottom of a natural stream. Assume a horizontal bottom with a current flowing in the x-direction, and the average horizontal velocity, $\bar{u}$, a constant for any given depth during the period of our investigation. If the bottom is considered to be hydrodynamically smooth, there will be a thin layer of laminar flow on the bottom even though the main flow be inherently turbulent. If the surface be hydrodynamically rough, then the turbulence effectively extends to the very bottom.

Flow in the laminar layer is characterized by the absence of random velocity fluctuations. In flow of this type, thin fluid layers are considered as moving parallel to each other in such a manner that the viscous stress τ exerted on a unit surface between these layers, the shear normal to that surface du/dz, and the coefficient of molecular viscosity μ, are related by Newton's fundamental equation

$$\tau=\mu\frac{du}{dz}. \tag{2}$$

An equation analogous to (2) may be written for the condition of turbulent flow:

$$\tau=A_z\frac{d\bar{u}}{dz} \tag{3}$$

where $\bar{u}$ is the average velocity and A_z is the value of the eddy viscosity A, applicable in the vertical direction. "A" is the exchange coefficient introduced by Schmidt (1917) as a result of the space exchange of momentum due to random fluctuation in turbulent flow. Both μ and A have the dimensions of $ML^{-1}T^{-1}$. It should be noted that the molecular viscosity is independent of the state of motion and is a characteristic property of the fluid. Observations show that the eddy viscosity is not a constant in natural currents, but depends upon the nature of the fluid movement. The eddy viscosity is many orders of magnitude larger than the molecular viscosity. For ocean currents A has been found to range between one and 1000 c.g.s. units (Sverdrup, *et al.*, 1942, p. 91).

Prandtl's Mischungsweg Theory leads to a definition of A_z as

$$A_z=\rho l^2\left|\frac{d\bar{u}}{dz}\right| \tag{4}$$

where l is the "mixing length," ρ is the density of the fluid, and $|d\bar{u}/dz|$ is the absolute value of the shear. Over a rough surface l must have a definite value, since turbulence extends to the very bottom, and for a fluid of indifferent stability may be written

$$l = k_0(z + z_0)$$

where k_0 is von Kármán's constant, and has been found to be equal to nearly 0.4. z is measured positive upwards, and z_0 is the roughness length and is related to the height of the roughness elements.

Combining equations (3) and (4) and considering the stress to be constant, one obtains the von Kármán-Prandtl equation for boundary currents over a rough surface:

$$\bar{u} = \frac{u_*}{k_0} \ln \frac{z + z_0}{z_0} \tag{5a}$$

where the factor u_* is equal to $\sqrt{\tau/\rho}$ and is called the "friction velocity." u_* has the dimensions of a velocity and is proportional to the average velocity $\bar{u}$. A coefficient γ is defined by

$$u_* = \gamma \bar{u}$$

and comparison with equation (5a) shows that γ is a function of the height at which $\bar{u}$ is measured above the bottom, and the degree of bottom roughness. Munk (1947) has reviewed the evidence which indicates that γ has a constant value over a rough surface, and decreases abruptly in magnitude at a critical wind speed below which the surface becomes smooth.

Equation (5a) gives the average velocity as a straight line function when plotted against the logarithm of $z + z_0$ and enables u_* and z_0 to be obtained from the slope of the velocity gradient and the z-axis intercept.

For a smooth surface equation (5a) is not applicable. The von Kármán-Prandtl equation for the velocity distribution in turbulent flow over a smooth surface is

$$\bar{u} = 5.5u_* + \frac{u_*}{k_0} \ln \frac{u_* z}{\nu} \tag{5b}$$

where ν is the kinematic viscosity, μ/ρ.

The stress in turbulent flow may be written (von Kármán, 1935) in the Reynolds form as

$$\tau = -\rho \overline{u'w'}$$

where u' and w' are the random velocity fluctuations in the x and z directions respectively. If it assumed that a correlation exists between u' and w', then the friction velocity is proportional to the mean of the absolute value of the velocity fluctuations. Further, if one assumes the turbulence to be isotropic, the friction velocity is equal to the mean of the absolute value of the velocity fluctuations

$$u_* = \sqrt{\frac{\tau}{\rho}} = \overline{|u'|} = \overline{|w'|}.$$

Roughness Criterion: From von Kármán's Law for flow over a smooth surface the thickness of the laminar layer may be written

$$\delta = \frac{\lambda \nu}{u_*} \tag{6}$$

where λ is a constant. It seems logical to assume that if the bottom is to be considered rough, the roughness elements (sand grains for example) must protrude above the laminar layer. This assumption suggests that some constant value of the ratio of bottom grain diameter d to the thickness of the laminar layer may be a criterion between rough and smooth bottoms. If one substitutes the value of δ from equation (6), the ratio d/δ may be written in the dimensionless form of a Reynolds number:

$$\frac{u_* d}{\nu} = \text{constant}.$$

The experiments of Nikuradse (1933), Fage (1933) and White (1940) indicate that the criterion between smooth and rough surfaces is a critical value of the ratio $u_* d/\nu$ such that:

$$\underset{\text{surface}}{\overset{\text{rough}}{\frac{u_* d}{\nu} > 3.5 >}} \underset{\text{surface}}{\overset{\text{smooth}}{\frac{u_* d}{\nu}}}.$$

One can therefore define a critical friction velocity $u_{*r} = 3.5\nu/d$ at which the character of the bottom changes from smooth to rough. This critical friction velocity will be referred to as the *roughness velocity.*

Settling Velocity: Newton (see Camp, 1946) proposed that a particle falling in a fluid will accelerate under the force of gravity until the "frictional drag" of the fluid approaches the value of the impelling force, after which a constant velocity c, called the settling velocity, will be reached. The frictional drag law was proposed on the assumption that the drag force F_D is proportional to the square of the velocity:

$$(7) \qquad F_D = C_D B \rho \frac{v^2}{2}$$

where C_D is a dimensionless number called the drag coefficient; B is the projected area of the particle in the direction of motion ($\pi d^2/4$ for spheres); and v is the relative velocity between the particle and the fluid.

Instead of being constant as Newton assumed, the drag coefficient is found to be a function of the Reynolds number $R = vd/\nu$.

When a small particle settles in a fluid it is acted upon by a force of gravity acting downward, and by a buoyant force (given by Archimede's principle) acting upward. The difference between these two forces is called the impelling force. For spherical particles it is equal to $1/6\pi g \rho' d^3$, where g is the acceleration due to gravity, and ρ' is the difference between the density of the particle and that of the fluid.

By equating the impelling force to the drag force, the settling velocity is found to be:

$$(8) \qquad c = \sqrt{\frac{4}{3} \frac{g}{C_D} \frac{\rho' d}{\rho}}.$$

In the region of viscous settling (diameters smaller than .18 mm) C_D is equal to $24/R$. Substituting this value of C_D in equation (8) gives Stokes Law for settling velocities of small particles.

$$c = \frac{1}{18} \frac{g}{\mu} \rho' d^2.$$

Particles with diameters greater than .18 mm have settling velocities sufficiently great to cause turbulence to develop, and the settling velocity no longer follows Stokes Law. The settling velocity (from Rubey, 1933) is plotted against the particle diameter in figure 2.

A spherical quartz grain of .18 mm diameter has a settling velocity in water of approximately 2 cm/sec. It is interesting to note that the Reynolds number of the grain at this velocity is equal to the critical value of the bottom roughness criterion

$$\frac{u_* d}{\nu} = 3.5.$$

THEORY OF SEDIMENT MOVEMENT

There appear to be three distinct methods of transportation of sediments by currents: *surface creep, saltation,* and *suspension.* The first is a bed load phenomenon in which the material slides or rolls along the bottom with a velocity less than that of the fluid current. Saltation is a jumping motion in which the particles momentarily bounce above the bottom and travel with the velocity of the fluid.

The forces associated with currents that cause a particle to move have been explained in the following ways (Rubey, 1938): (1) by the impact or momentum of the water which strikes it, (2) by frictional drag on its surface, and (3) by differences in pressure induced by differences in current velocity.

The impact principle (giving the 6th

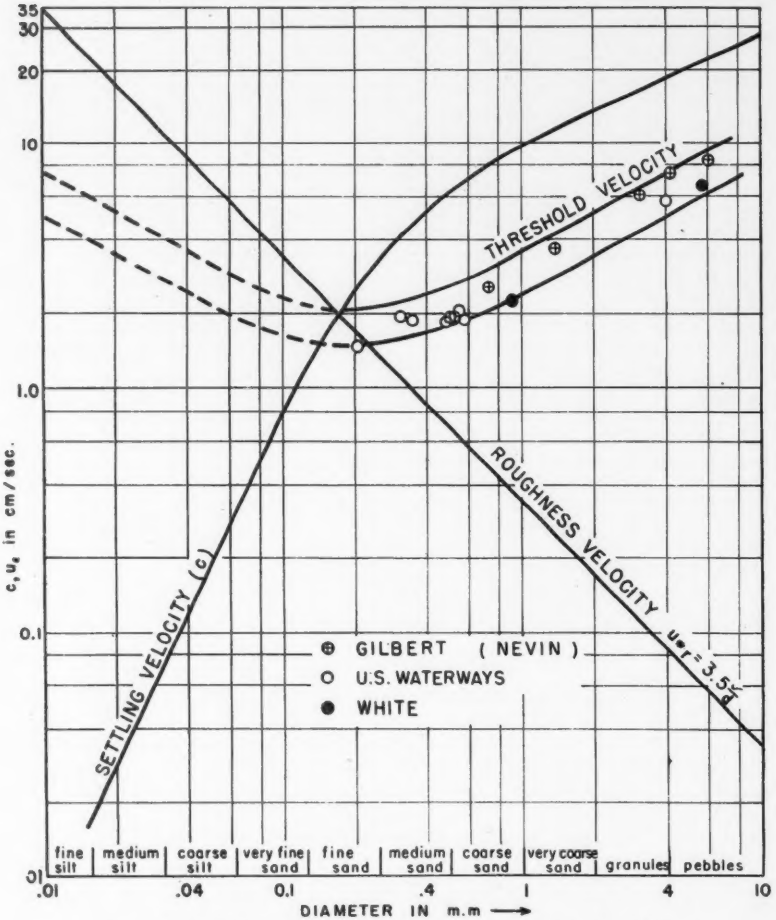

FIG. 2.—Relation of grain diameter to settling velocity, threshold velocity and roughness velocity.

power law) is difficult to evaluate in terms of bottom currents, since it necessitates measuring the current at the boundary surface. Consideration of frictional drag leads to the conclusion that a critical tractive force exists which is important to sediment movement.

Surface Creep: Experiments by Shields (1936), White (1940), Bagnold (1942) and others indicate that movement will not take place until a critical drag force that is a function of grain size is exceeded. The critical drag force f_c on a spherical particle may be written as

$$f_c = \alpha g \rho' d^3 \tan \phi$$

where α is a factor depending upon the type of flow, and ϕ is the angle of repose of the grains. If the sand is of uniform diameter d, the number of grains per unit area that are exposed to the fluid drag are p/d^2, where p is the packing coefficient and is defined as d^2 times the number of exposed grains per unit area. The critical drag force per unit area τ_c is equal to $f_c p/d^2$, or

$$\tau_c = p\alpha g\rho' d \tan \phi. \tag{9}$$

The factor α is a function of the roughness criterion $u_* d/\nu$, and experiments show it to have a value of .25 for values of $u_* d/\nu$ greater than 3.5 (White, 1940). The critical drag may be expressed in terms of a critical frictional velocity u_{*c}

$$u_{*c} = \sqrt{\frac{\tau_c}{\rho}}.$$

This critical value of u_* at which sand movement starts, owing to the drag force of the fluid, is called the "threshold velocity" (Bagnold, 1942, p. 86).

We have dealt in general terms with a critical drag force or its equivalent threshold velocity which may be computed from the average velocity profile (see equation 5a). However, it should be remembered that if even momentarily the value of u_{*c} is exceeded, movement will result. Thus we are interested in the instantaneous velocity u, consisting of the average velocity $\bar{u}$, plus the velocity related to random fluctuation u' (see equation 1). Kalinske (1943) shows that the instantaneous drag force may be many times that computed from the average velocity.

A broad curve including some of the experimental values of the threshold velocity for various grain diameters is shown by parallel solid lines in figure 2. These threshold velocities were computed from the experimental critical drag forces obtained by White (1940), the U.S. Waterways Experiment Station, and from Nevin's (1946) analysis of data obtained by Gilbert. Since it is not always practical to measure velocity profiles in shallow water flume experiments, the critical drag forces in the above mentioned experiments were computed from du Boys' equation. This may be written in c.g.s. units as

$$\tau = g\rho Di$$

where D is the depth of flow and i is the slope of the bed. The bed slope is assumed to be equal to the slope of the water surface. The threshold velocity gradually decreases with decreasing grain diameter until a diameter of approximately .18 mm is reached. At this point the value of $u_* d/\nu$ becomes less than 3.5, the bottom becomes effectively smooth, and individual grains cease to shed small eddies. The drag, instead of being carried by a few grains, is more evenly distributed over the whole surface. As a result of these changes the velocity necessary to start surface movement increases as the grain diameter decreases below approximately .18 mm. This startling fact was experimentally demonstrated (Bagnold, 1937) by blowing a stream of air over a loosely scattered layer of fine Portland cement powder. It was found that no movement occurred even when the wind was strong enough to move pebbles 4.6 mm in diameter.

It should be noted that while the relationship $u_* = \gamma \bar{u}$ holds for a smooth surface as well as for a rough surface, the value of γ is not the same in both cases. The use of a constant value of γ in computing threshold velocity from the average stream velocity may not give accurate values when the bottom changes from rough to smooth. However, the stream velocity required to move the finer material does increase with decreasing grain size for water as well as wind transported sediments (Hjulström, 1939). This fact is illustrated in figure 2 by dotted lines representing the *apparent* threshold velocity. The data for this section of the curve were obtained from average stream velocities tabulated by Hjulström (1935).

Saltation: As soon as movement of sand has been initiated, some of the particles may be lifted off the bed into the fluid stream. Once in the fluid stream, the particles acquire the velocity of the fluid and travel with it until gravity forces them back to the surface of the bed. The particle may "bounce," or strike another particle, causing it to leave the surface. Jeffreys (1929) has shown that the initial rise of particles from a bed may be partly explained on a basis of Bernoulli's principle giving a resultant upward force in the presence of a steep velocity gradient. Here again, u_* may become an important parameter since it is related to the velocity gradient.

Kalinske (1943) believes that saltation is not important in water, although it has been observed by Gilbert (1914), Bagnold (1942) and others. Certainly it loses its significance for fine grain sediments.

Suspension: Transportation of sediment in suspension is a problem both in diffusion and advection. A qualitative relationship for material in suspension under conditions of equilibrium was first developed by Schmidt (1925) for air and later applied to silt in water by O'Brien (1933). Schmidt assumed that the mass of sediment particles carried downward by gravitational settling must equal the amount carried upward by turbulent motion. This relationship may be written in terms of the distance above the bottom z and the eddy viscosity A_z as

$$(12a) \qquad \ln \frac{S_z}{S_a} = -c \int_a^z \frac{dz}{A_z/\rho}$$

or, in terms of the velocities at these elevations and the friction velocity u_* as

$$(12b) \qquad \ln \frac{S_z}{S_a} = -c \int_a^z \frac{d\bar{u}}{u_*^2}$$

where S_z/S_a is the ratio of the concentration of sediment having a settling velocity c, at height z to that at a constant height "a."

The theory of the vertical distribution of sediment in suspension is dependent on a concentration gradient. It should be noted that under similar flow conditions, the ratio of the sediment concentrations is proportional to e^{-c}, where e is the base of natural logarithms. Thus fine material has a more uniform distribution with depth, and hence a smaller concentration gradient than coarse material. If the height "a" is considered to be infinitely close to the bottom, and the turbulence is sufficient to cause suspension, then the concentrations of the various grain sizes at height "a" are proportional to the amount of each grain size on the surface of the bottom.

The significance of the suspension equations may be better understood by further development of equation (12b). Integrating (12b) and using (5a) and (5b) we obtain

$$(13) \qquad \log_{10} \frac{S_z}{S_a} = -\frac{1}{k_0} \frac{c}{u_*} \log_{10} \frac{z+z_0}{a+z_0} .$$

Here $z_0 = 0$ corresponds to flow over a smooth surface.

If we assume that the particles in suspension are spherical, of diameter d, and of the same specific gravity, then the settling velocity of the particles (equation 8) is a function of d only: $c = f(d)$. Setting $u_* = nc$, where n is any positive number, we obtain

$$(14a) \qquad \log_{10} \frac{S_z}{S_a} = -\frac{1}{nk_0} \log_{10} \frac{z+z_0}{a+z_0} .$$

Equation (13) tells that if we select fixed values of z and "a," any curve for which the ratio u_*/c equals a constant is a curve along which the concentration ratio S_z/S_a is also a constant. In a coordinate system u_* vs. d we can therefore construct a family of curves for constant values of the concentration ratio. For the sake of simplicity we may select

$$\frac{z+z_0}{a+z_0} \doteq \frac{z}{a} = 10$$

and obtain

$$(14b)\quad \log_{10}\frac{S_z}{S_a}=-\frac{1}{nk_0}\qquad (\text{for } z=10a).$$

It will be noted from equation (14b) that when u_* equals c, $n=1$ and

$$\log_{10}\frac{S_z}{S_a}=-2.5\qquad (\text{for } z=10a)$$

where k_0 is equal to 0.4.

This value of the logarithm of the concentration ratio will be used as the limiting value for which suspension is significant. In effect, this means that if suspension is to exist to a significant degree, then u_* must be equal to, or greater than, c; and under our arbitrary selection of elevations $z=10a$, the logarithm of the concentration ratio must be equal to, or greater than, -2.5.

This selection is in agreement with an analysis of experimental data by Lane and Kalinske (1939, also Kalinske, 1941) which indicated that a value of unity for n, the ratio of the friction velocity u_* to the settling velocity c, is a criterion for the beginning of suspension. This criterion can also be justified on the basis that, since u_* is equal to the mean of the absolute value of the vertical turbulent velocity fluctuations $\overline{|w'|}$, it should be expected that a significant degree of suspension can exist only if the settling velocity is less than $\overline{|w'|}$.

A number of curves of constant concentration ratios are shown in figure 3. The limiting value of -2.5 for which suspension is significant is shown as a heavy line. This curve for which $u_*=c$ may also be interpreted as a plot of the settling velocity against the particle diameter d. It corresponds to the settling velocity curve in figure 2.

In the application of suspension to sorting and skewness frequent reference will be made to the settling velocity curve in figure 2 and to the role of settling velocity as a criterion for suspension. However, it should be remembered that the settling velocity curve is but one of the family of curves illustrated in figure 3, its significance being that it is the particular curve for which the threshold velocity is equal to the settling velocity.

APPLICATIONS OF HYDRODYNAMIC THEORY

Some of the many complex factors that enter into the problem of the interrelationships of sorting, skewness, and median diameter have been discussed above. It is necessary to consider the effect of these factors in both the transportation and deposition of sediments. In the laboratory, experiments have been performed under controlled conditions on various phases of the sedimentation problem, but in nature not only are the conditions far from being controlled, but the relative importance of the different factors is not clearly understood.

In figure 2, three parameters, the settling velocity, the roughness velocity and the threshold velocity, are plotted on double log paper with the velocity components c and u_* as the ordinate and the grain diameter in millimeters as the abscissa. These relationships are for spherical grains with a specific gravity of 2.65 in water at 20° centigrade.[1] It should be remembered that this graph is drawn for homogeneous grains of one size, and that studies of sorting and skewness involve aggregates of two or more grain

[1] A preliminary inspection, not based on experimental data, indicates that the three curves in figure 2 will retain their general relationship for changing particle specific gravity (water being the fluid media). Since the roughness criterion is independent of the grain specific gravity, it will remain a constant. Material with a specific gravity greater than 2.65 will have a higher settling and threshold velocity, so that the three curves should intersect on the roughness velocity at a smaller grain diameter and a higher value of u_* and c. Theoretical calculations with grains of magnetite (specific gravity 5.17) gave an intersection of the curves in the very fine sand grade. For specific gravity less than 2.65, the three curves should intersect at a greater grain diameter and a smaller value of u_* and c.

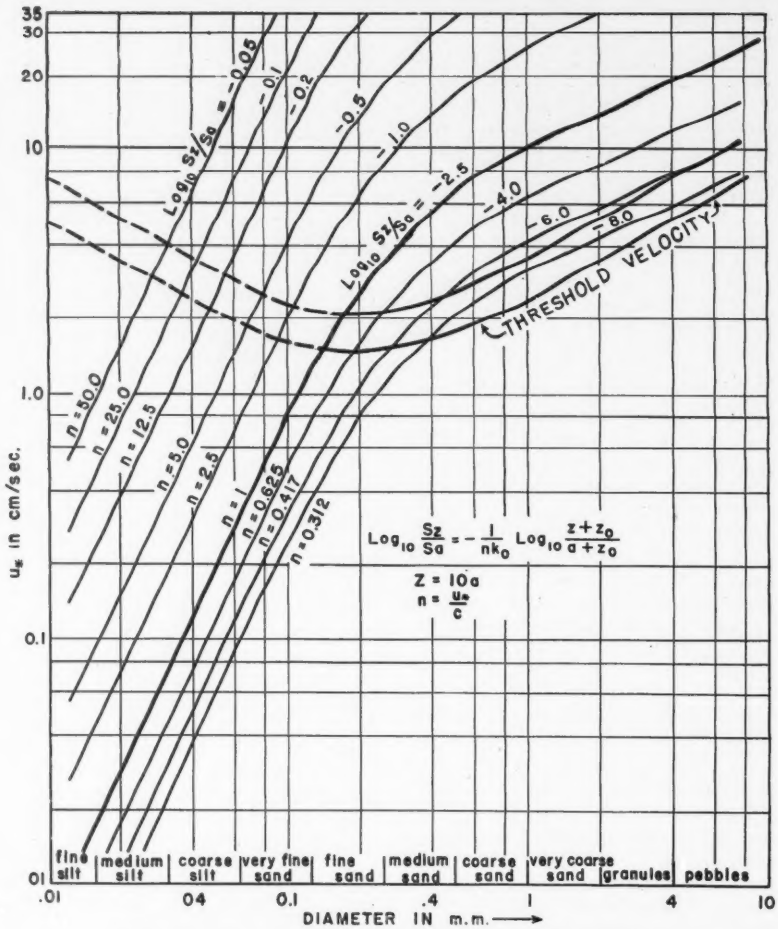

FIG. 3.—Relation of the concentration ratio of suspended material to the threshold velocity.

sizes. While the use of such a graph tends to over-simplify the true picture, it is hoped that it will serve to illustrate some of the fundamental characteristics of the different grain sizes.

An examination of figure 2 will illustrate some of these characteristics:

1. The Wentworth grade classification of "fine sand" with an average diameter of .18 mm is the most easily moved, and the value of u_* necessary to cause initial movement and to cause roughness is numerically equal to the settling velocity. This indicates that grains of this size may be transported at relatively low velocities by surface creep, saltation and in suspension.

2. The threshold velocity for grain

sizes greater than .18 mm increases as the grain diameter increases. For these grain sizes the threshold velocity is greater than the roughness velocity and less than the settling velocity, indicating that the bottom is rough before movement takes place; and that when movement is initiated, it will be in the form of rolling and sliding rather than suspension.

3. The threshold velocity for grain diameters less than .18 mm increases with decreasing grain size. As previously mentioned this is probably due to the fact that the bottom is smooth and the drag is evenly distributed over the entire area rather than being applied to a few of the more exposed grains. Also, for these small grains the threshold velocity is much greater than the settling velocity. This indicates that if once put in suspension they will tend to be transported in suspension rather than by surface creep.

It is not clearly understood how material on a smooth bottom is put is suspension. Suspension over a hydrodynamically smooth bottom may be due to the presence of a lifting force as proposed by Jeffreys' (1929) and strengthened by White's (1940, p. 333) observations, or it may be due to increased bottom roughness by the formation of ripples after movement has beeen initiated.

Sorting: Russell (1939) divides sorting action into two types: "local" sorting involving the assortment of particles at a particular locality or site of deposition; and "progressive" sorting consisting of an assortment in the direction of transportation. Russell states: "The most important factors involved in both types of sorting appear to be the size, shape, and specific gravity of the particles, and the velocity, degree of turbulence, viscosity, and specific gravity of the transporting agent." It is also necessary to consider the methods of transportation and whether the sorting takes place during transportation or deposition or both. In this paper our discussion will be limited to spherical particles with a specific gravity near 2.65.

It would seem that the degree of sorting would be a function of the ability of the fluid to sort out one grain size from another. Grain sizes near .18 mm are hydrodynamically unique from all other sizes in that: (1) they are moved by weaker currents than grains smaller or larger than themselves; (2) once moved they do not have as great a tendency to go into suspension as do smaller grains; and (3) they are more readily carried into suspension than larger material. Thus where friction velocity does not greatly exceed the threshold velocity, fine sand marks the transition zone between transportation in suspension and transportation by surface creep. An inspection of figure 2 indicates that a current with a friction velocity that gradually decreases in the direction of transport will move bottom material near .18 mm a greater percentage of the time than any other diameter. Since both coarse and fine materials are more difficult to move and since very fine material is readily carried into suspension, bottom sediment in the process of transportation tends to become progressively better sorted as its median diameter nears .18 mm.

Sediments with median diameters either larger or smaller than .18 mm tend to be more poorly sorted; the tendency toward poor sorting being more pronounced for fine sediments. This may be accounted for by the fact that the fluid does not as readily differentiate between the smaller diameters. The fine sediments are more closely packed and the drag is evenly distributed over them. For a homogeneous size distribution the apparent threshold velocity increases with decreasing grain size below that of fine sand, however it seems probable that for an aggregate of fine sizes the threshold velocity is effectively constant for diameters less than .18 mm, so that different diameters of fine sediments will tend to move by surface creep as a group. Once in suspension, they will be completely removed to an area of much less current since the settling velocity is small compared with the threshold velocity.

In the absence of movement by surface creep, the great differences in settling

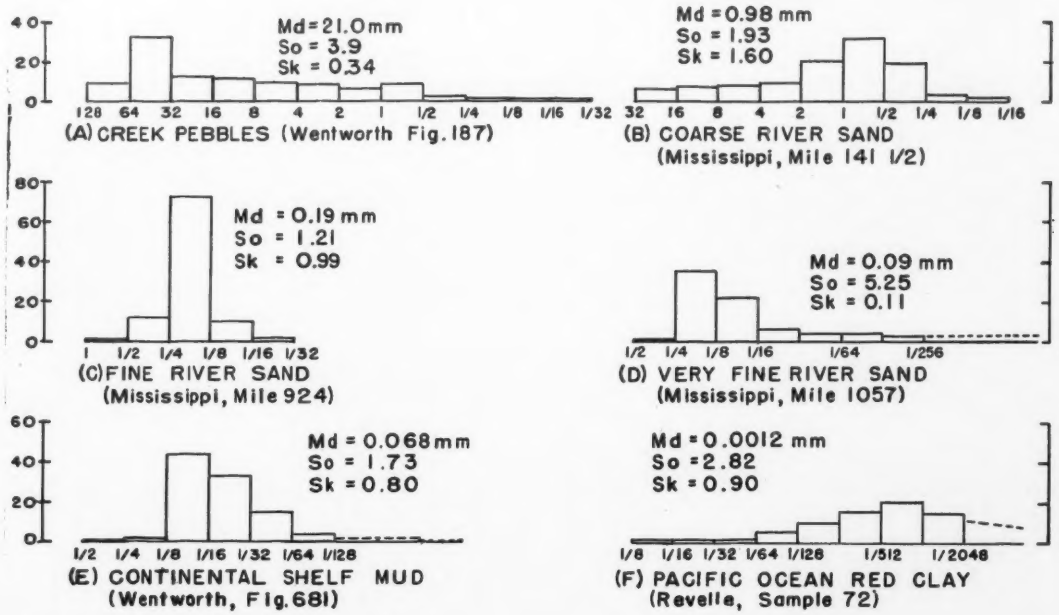
Md=21.0mm
So = 3.9
Sk = 0.34
128 64 32 16 8 4 2 1 1/2 1/4 1/8 1/16 1/32
(A) CREEK PEBBLES (Wentworth Fig. 187)
Md = 0.98 mm
So = 1.93
Sk = 1.60
32 16 8 4 2 1 1/2 1/4 1/8 1/16
(B) COARSE RIVER SAND
(Mississippi, Mile 141 1/2)
Md = 0.19 mm
So = 1.21
Sk = 0.99
1 1/2 1/4 1/8 1/16 1/32
(C) FINE RIVER SAND
(Mississippi, Mile 924)
Md = 0.09 mm
So = 5.25
Sk = 0.11
1/2 1/4 1/8 1/16 1/64 1/256
(D) VERY FINE RIVER SAND
(Mississippi, Mile 1057)
Md = 0.068mm
So = 1.73
Sk = 0.80
1/2 1/4 1/8 1/16 1/32 1/64 1/128
(E) CONTINENTAL SHELF MUD
(Wentworth, Fig. 681)
Md = 0.0012 mm
So = 2.82
Sk = 0.90
1/8 1/16 1/32 1/64 1/128 1/512 1/2048
(F) PACIFIC OCEAN RED CLAY
(Revelle, Sample 72)

FIG. 4.—Histograms of typical sediment types
(B), (C), and (D) are from the U. S. Waterways Experiment Station (1935); the numbers refer to miles below Cairo, Illinois.

velocities of very fine sediments act over large lateral distances as an effective sorting agent. For this reason very fine sediments, such as deep basin muds and clays derived only from the deposition of suspended material, may be better sorted than sediments with median diameters intermediate between them and fine sand.

Shallow Water Sediments Under Unidirectional Flow: Consider an hypothetical case in which the average velocity $\bar{u}$ and the random fluctuations u' near the bottom of a river or ocean current gradually decrease in magnitude in a downstream direction, and the initial upstream current causes a stress on the bottom sufficient to move all bottom material. Consider the supply of material to be continuous and to consist of all sediment grades from clays to cobbles. Under these flow conditions there will be a decrease in the threshold velocity in a downstream direction, and hence a gradual decrease in the median diameter[1] of the bottom samples. For a given size material this decrease in median diameter will depend upon whether the stream is in a condition of "degradation," "at grade," or "aggradation."

Consider the character of bottom samples obtained at various points downstream from the source material, when transported under the above flow conditions. Near its source where the stream is actively degrading its channel, the high values of friction velocity would cause the fine material including sand to be mostly in suspension. However, to maintain a suspended load, a portion of the fine material will be on the bottom. The bottom load at this point would consist predominantly of coarse material with decreasing amounts of fine. Since the sample is near the source and the friction velocity exceeds the threshold velocity for all but the coarsest material, it is to be expected that the sample will not be as well sorted as material farther downstream. Also, since it consists of coarse material with decreasing amounts of fine, the histogram of the sample will tend to be skewed toward the fine fractions. In other words, the median would be finer than the mode (see fig. 4a). It seems probable that the common finding of skewness values less than one in very coarse samples is due to a lag condition of this type.

Consider a bottom sample from a point farther downstream where the average friction velocity fluctuates between the threshold velocity for very coarse sand and granules (see fig. 2). Under these conditions it would seem that the bottom material transported by surface creep would consist predominantly of coarse and very coarse sand with decreasing amounts of granules and pebbles. There would also be finer grades present on the bottom, their amounts decreasing with decreasing diameter. This does not imply that the stream is transporting more coarse than fine material. The presence of coarse material would probably produce sufficient bottom turbulence to cause fine material to be placed in suspension. Since finer diameter materials have lower settling velocities, less material per unit area is required on the surface of the bottom to support a dense suspended load (see equation 13). It seems probable that medium sand would be transported both by saltation and in suspension; fine sand would be transported predominantly in suspension but partly by saltation; and very fine sand, silts and clays would be transported in suspension.

A bottom sample from the point where the friction velocity fluctuates between the threshold velocity for very coarse sand and granules would be better sorted than samples farther upstream because the flow characteristics of the stream that tend to sort material have acted over a greater time and distance. Also, the fine material is still largely transported in suspension, and the quantity of material transported by surface creep is inversely

[1] The decrease in grain size due to abrasion may be an important factor, but due to the many variables it is not practical to consider here.

proportional to the particle diameter. These factors give a more homogeneous size distribution of bottom material (see fig. 4b).

It might be expected that the histogram of this sample would be skewed toward the coarse grades, and consist predominantly of coarse and very coarse sand with decreasing amounts of granules and pebbles. However, Gilbert (1914), Chang (1939) and others have shown that once movement is initiated, larger particles tend to move faster and stop less frequently (due to greater momentum and the fact that their larger diameter places them in a faster field of flow) than finer particles. Mississippi River bottom samples collected by the U. S. Waterways Experiment Station (1935) indicate that samples with median diameters near 1.0 mm are predominantly skewed towards the coarse grades (fig. 4b), while those with median diameters between 1.0 and .18 mm are nearly symmetrical or have skewness values slightly less than one.

Next, consider a bottom sample from a downstream point where the average friction velocity fluctuates between the threshold velocity for fine and coarse sand. Bottom roughness would produce sufficient turbulence to cause suspension of fine material, including some fine sand. For reasons previously mentioned, the fine sand would be most easily moved; the coarser and finer grades moved by surface creep would tend to lag behind. For these reasons the bottom sample would have a median diameter near .18 mm, it would be well sorted, and have a symmetrical size distribution (fig. 4c).

As a final hypothetical case, consider the character of a bottom sample still farther downstream from the above points where the stream has a friction velocity near, but not exceeding, the threshold velocity of fine sand. Therefore, material coarser than fine sand would be almost entirely absent. The bottom material transported by surface creep would consist predominantly of fine sand with decreasing amounts of finer material The presence of fine sand would produce sufficient bottom turbulence to cause finer material to be in suspenion. However, the decrease in friction velocity would require a large percentage of fine material per unit area on the surface of the bottom to maintain a suspended load. Thus the bottom sample would consist mainly of fine sand with gradually decreasing amounts of finer material. The sample would have a median diameter less than .18 mm, it would not be as well sorted as slightly coarser material, and the histogram of the sample should show a pronounced skewness toward the fine fractions (fig. 4d).

Sediments found under flow conditions similar to those postulated above are in agreement with the hypothetical bottom samples. This agreement is strengthened by an inspection of the 615 Mississippi River bottom samples collected from Cairo, Illinois, to the mouth of the river by the U. S. Waterways Experiment Station (1935). Since the U. S. Waterways' samples ranged in location from mid-channel to bars, and were often influenced by sediments from tributary streams, it is to be expected that individual samples will vary widely in character. However, their general tendencies are significant. The sediments have a sorting coefficient near 3 just below Cairo Illinois, where the median diameter falls in the coarse sand grade. The median diameter decreases downstream and the sorting improves. A sorting of 1.2 is obtained above New Orleans where the median diameter is .18 mm. Below this point the sorting coefficient has a value greater than 5 for bottom samples with median diameters near .09 mm. The bottom samples near Cairo are predominantly skewed toward the coarse grades. The symmetry of the sample improves in a downstream direction. Near New Orleans where the median diameter is .18 mm, the samples have a symmetrical size distribution. As the median diameter becomes finer below New Orleans, the histogram of the bottom samples show a marked skewness toward the fine grades. A graphic representation of the distribu-

tion of sediment sizes in the Mississippi River samples, averaged by 25 mile intervals, is shown in figure 5.

A preliminary examination of that portion of the 828 typical sediment samples collected by Wentworth (1932) that relate to the present discussion show that samples having median diameters near .18 mm are predominantly well sorted and symmetrical. Samples having median diameters less than .18 mm are predominantly poorly sorted and skewed toward the fine grains. The coarser median diameter samples are more poorly sorted than fine sand samples and are skewed toward both the coarse and the fine. Coarse sand and granules from rivers and creeks were predominantly skewed toward the coarse grades. Pebbles and gravels from rivers and creeks were predominantly skewed toward the fine grades.

Sediments Derived from Suspension: In discussing fine sediments it should be noted that the colloidal fractions (less than 1 micron) of a sample are not necessarily subject to the hydrodynamic relations discussed in this paper. Furthermore, Gripenberg (1934) and Rouse (1939) point out that material less than 15 microns in diameter will often flocculate, and that mechanical analysis of deflocculated samples may yield histograms which bear little relation to the actual material in suspension.

Rouse (1939), in an analysis of the effect of turbulence on transportation of sediments in suspension, showed that with increasing height above the bottom, the sediment in suspension becomes more poorly sorted and its median diameter becomes finer. From these observations and the fact that bottom-transported fine sediments are predominantly skewed toward the fine, one would expect that pelagic sediments derived only from suspension (i.e., deep basin sediments) would tend to be nearly symmetrical or skewed slightly toward the coarser grades. As mentioned under sorting, very fine sediments may be better sorted than sediments with median diameters intermediate between them and fine sand. This ideal sample is frequently complicated in nature by the settling of organic material (calcareous and siliceous ooze) from the euphotic zone. Organic samples of this type are not subject to the same conditions as the inorganic since they have travelled relatively short horizontal distances. The fact that they are not subjected to the same transportation conditions is illustrated by the frequent double maxima in deep sea sediments. Revelle (1944) showed that double maxima were frequently due to organic material.

Deep basin mud and clay samples were obtained in the Pacific by the Carnegie Expedition (Revelle, 1944) and in the North Atlantic by the Meteor Expedition (Correns, 1939). Neglecting samples with double maxima, these deep sea sediments are characterized by a medium degree of sorting, and in general are either skewed toward the coarse grade sizes or have skewness values near one (figure 4f). Gripenberg (1934, p. 203) discusses some of the factors influencing the size distributions of bottom samples from the North Baltic Sea. Many of the histograms of her samples are skewed toward the coarse grades.

Beach Sediments: It is well known that beach sands are among the best sorted sediments. They are subjected to a progressive sorting when transported by longshore currents, and to a very effective local sorting by the oscillatory motion of the waves. Observations of southern California beach sands indicate that their sorting coefficients bear the same general relationship to median diameter as sediments from other environments, the difference being mainly in the degree of sorting (fig. 1).

Grant (1943) points out that theoretical studies and observations of oscillatory wave phenomena prove the existence of a shoreward mass transport of water, and that there is evidence of a differential in the velocity of the offshore and onshore orbital movement in the surf zone. LaFond (1939) shows that a

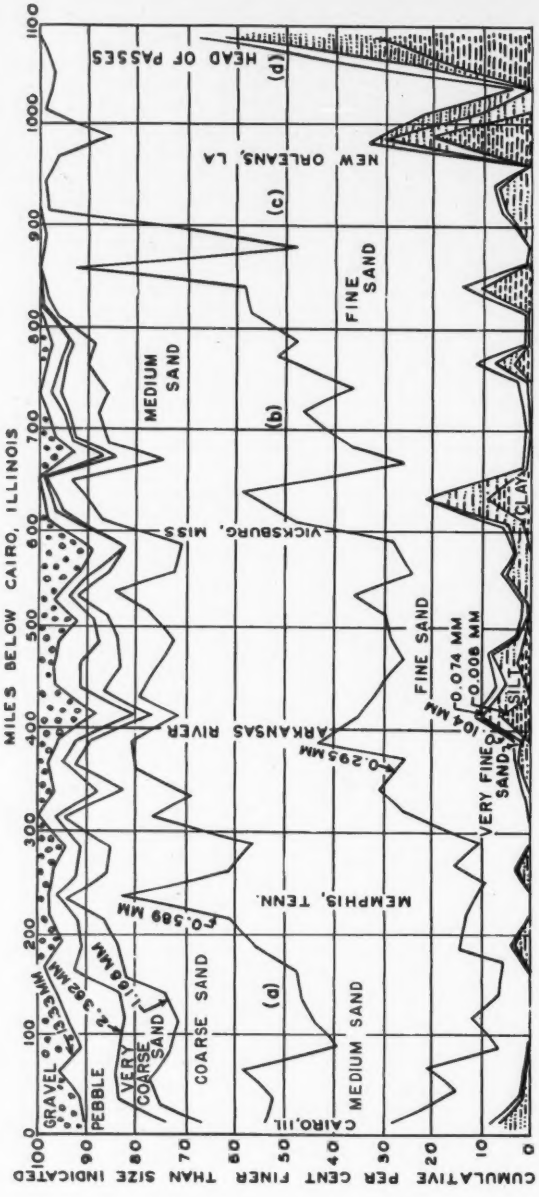

FIG. 5.—Variation in composition of Mississippi river bed materials (from the U. S. Waterways Experiment Station, 1935). Statistical parameters of samples at localities indicated by letters are:

	Md(mm)	*So*	*Sk*
(a)	0.98	1.93	1.60
(b)	0.42	1.41	1.18
(c)	0.19	1.21	0.99
(d)	0.09	5.25	0.11

direct relationship exists between the fluctuations in the amount of sand on the beaches and the variations in wave size and type.

The orbital or particle motion associated with shallow water surface waves, relative to a fixed point, is thought to consist of a slow seaward velocity under the wave trough, followed by a rapid shoreward surge as the crest passes over the point. Adaptation of the solitary wave theory to the surf zone (Munk, 1949) gives quantitative values to these relationships and, if valid, indicates that the differential between the crest and trough velocities decreases as the wave height and frequency increase.

The shoreward mass transport of water apparently causes a rise in the sea surface near shore. This "piling up" of water must be compensated by a seaward return, part of which is known to be in the form of rip currents (Shepard, *et al.*, 1941). The solitary wave theory requires that the shoreward mass transport attain a maximum at the water surface and decrease with depth. Munk suggests that, in addition to rip currents, this transport is compensated by a uniform seaward return flow. This would result in a net shoreward drift near the water surface and a net seaward drift near the bottom.[1] The intensity of the seaward drift along the bottom would be a function of wave height and period, and may be negligible for low amplitude long period waves.

If one assumes that all these facts are valid when applied to the sorting of beach sediments, it would appear that sands, and particularly fine sands, would be the most transient beach material. Under waves of medium intensity, fine sand would be moved by both crest and trough orbital velocities, thus tending to give fine sand a zero net transport. Under the same conditions coarse material would be moved only by the rapid surge associated with the wave crest, and hence would move shoreward. Fine material would have a greater tendency to be in suspension and would be transported seaward by rip currents and any net bottom drift.

Under less intense wave conditions, sand size material would have a net shoreward transport where as very fine material would still have a tendency to be transported seaward.

Under conditions of high amplitude short period waves, sand size material would be transported both by crest and trough velocities. However, the increased friction velocity would cause sand size material to be in suspension, and thus have a net seaward transport. This analysis is an amplification of suggestions made by Grant (1943) and by Munk (1949). It should be noted that there will always be a size material for which the net transport will be zero, and if such material is available, it will become the predominant foreshore beach material. It will be locally well sorted, since finer material will be transported seaward and slightly coarser material will be deposited on the beach. The presence of cobbles at the base of most sand beaches may be attributed to the fact that the largest material moved will be transported shoreward.

CONCLUSIONS

The general relationship of sorting coefficient to median diameter appears to be similar for all water environments, the difference being mainly in the degree of sorting. Sediments with median diameters near the grade of fine sand are the best sorted; sediments coarser and finer are more poorly sorted.

The relationship of skewness to median diameter is more complex (fig. 1). Inspection of Mississippi River bottom samples from a point well above Cairo to

[1] If such a drift exists, its magnitude does not exceed a few centimeters per second, and thus should not be referred to as "undertow." Computations based on the solitary wave theory (Munk, 1948) indicate that a 4 foot high, 8 second period wave in water 15 feet deep would have a 119 cm/sec. orbital crest velocity, a 67 cm/sec. orbital trough velocity (seaward), and a net seaward bottom drift of 1.5 cm/sec.

the mouth of the river indicate that plots of skewness coefficient against median diameter go through several cycles (fig. 4).

The relationship of settling velocity, threshold velocity and roughness velocity to the diameter of the grain size (fig. 2) may serve to explain some of the observed correlations between sorting, skewness and median diameters in actual sediment samples. The curves in figure 2 are supported by theory and experimental observations. However, it should be remembered that the hypothetical examples given above were limited in scope and were discussed for rigidly controlled conditions. In nature the amount and type of sediment source, the degree of turbulence and the magnitude of the currents and their fluctuations, the distance the material has been transported, reduction of grain size by abrasion, and the time intervals involved, are all variables that must be considered.

Krumbein (1939) illustrates with isopleth maps of the sediment parameters of Barataria Bay "that the characteristics of the sediments reflect the general conditions of their formation. . . ." It is hoped that in the future more detailed analysis of the interrelationship of the statistical sedimentary parameters, sorting, skewness and median diameter, to the hydrodynamic conditions of transportation will be made, so that these sedimentary parameters will become a more definite clue to the processes under which the sediment was transported and deposited.

NOTATION AND DIMENSIONS

(All units are in the c.g.s. system; M is mass in grams, L is length in centimeters, T is time in seconds.)

Symbol	*Description*	*Dimension*
a	Reference level used in equations 12, 13 and 14	L
A	Eddy viscosity	$ML^{-1}T^{-1}$
B	Projected area of a particle in the direction of motion	L^2
c	Settling velocity	LT^{-1}
C_D	Drag coefficient	Dimensionless
d	Particle diameter; centimeters for all equations, millimeters for graphs	L
e	Base of natural logarithms	Dimensionless
f_s	Critical tractive force on a spherical particle	MLT^{-2}
F_D	Frictional drag force on a moving particle	MLT^{-2}
g	Acceleration due to gravity	LT^{-2}
i	Slope; value in radians of the angle between the water surface and a horizontal plane	Dimensionless
k_0	von Kármán's flow constant, equals approximately 0.4	Dimensionless
l	Prandtl's mixing length	L
ln	Natural logarithm	
Md	Geometric median diameter; millimeters	L
n	Positive number equal to u_*/c	Dimensionless
p	Packing coefficient	Dimensionless
Q_1, Q_3	First and third quartile measures; millimeters	L
R	Reynolds number; $\frac{vd}{\nu}$	Dimensionless
S	Concentration of suspended particles expressed as mass per unit volume of water	ML^{-3}
Sk	Skewness coefficient; $\frac{Q_1Q_3}{Md^2}$	Dimensionless
So	Sorting coefficient; $\sqrt{\frac{Q_1}{Q_3}}$	Dimensionless
u	Instantaneous velocity in the x-direction; positive downstream	LT^{-1}

$\bar{u}$	Mean downstream velocity	LT^{-1}
u'	Velocity of random fluctuation related to turbulence in a downstream direction	LT^{-1}
u_*	Friction velocity; $\sqrt{\frac{\tau}{\rho}}$ u_{*c} is the threshold velocity necessary to move a given size particle u_{*r} is the roughness velocity	LT^{-1}
v	Velocity of a particle relative to the fluid	LT^{-1}
w'	Random fluctuation related to turbulence in the z-direction; positive upward	LT^{-1}
x	Horizontal axis; positive downstream	L
z	Vertical axis; positive upward	L
z_0	Roughness length, related to the height of the roughness elements	L
α	Factor depending upon the type of flow	Dimensionless
γ	Factor depending upon the height at which $\bar{u}$ is measured above the bottom and the degree of bottom roughness; γ^2 is commonly referred to as the resistance coefficient (Munk, 1948)	Dimensionless
δ	Thickness of the laminar boundary layer	L
λ	Flow constant (usually defined as $\lambda + \frac{1}{k_0} \ln k_0 = 5.5$)	Dimensionless
μ	Molecular viscosity	$ML^{-1}T^{-1}$
ν	Kinematic viscosity; μ/ρ	L^2T^{-1}
ρ	Mass density of the fluid	ML^{-3}
ρ'	Difference between mass density of the particle and that of the fluid	ML^{-3}
τ	Intensity of fluid shear per unit area; tractive force per unit area	$ML^{-1}T^{-2}$
τ_c	Critical tractive force per unit area exerted on the surface of the bottom by the fluid	$ML^{-1}T^{-2}$
ϕ	Angle of repose (degrees)	Dimensionless

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FIG. 1, AUGUST 1949

SYNTHESIS OF SAND MIXTURES*

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ABSTRACT

A method is offered for synthesizing sand mixtures with a normal probability size-frequency distribution, and any desired mean and sorting. Tables and a graph permit rapid use of the method for certain phi standard deviations and for any phi mean diameter. Graphical and mathematical solutions are also presented so that any normal sand may be mixed.

NOTATION

C_s = coefficient of sorting
M = mean diameter
Md = median diameter
M_ϕ = phi mean diameter
$M_{\phi n}$ = nominal phi mean diameter
Q_1 = larger quartile
Q_3 = smaller quartile
$\phi = -\log_2 n$, where n is the grain size in millimeters. If $M\phi_1 = M\phi_2 + \phi$, then $M_1 = 2M_2$
σ_ϕ = phi standard deviation

INTRODUCTION

Sediments are analyzed every day but they are seldom synthesized. The author has mixed sands in order to measure the effect of sorting on the transportation of bed-load. Krumbein and Monk (1942) have mixed them previously to determine the influence of sorting on permeability. Some solutions to the problem of synthesizing a sand mixture with a natural size-frequency distribution and given characteristics are presented here. The general procedure employed is to sieve one or more natural sands in order to obtain a number of size fractions. These fractions may be combined in proportions determined by the graph and tables included in this paper to synthesize a sand.

The author is indebted to the Woods Hole Oceanographic Institution for a summer fellowship which made this study possible. Mr. Joseph Rominger and Mr. Henry C. Stetson have been kind enough to read the manuscript. Mr. Rominger has corrected several errors and pointed out the paper by Krumbein and Monk.

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SIZE-FREQUENCY DISTRIBUTIONS IN NATURAL SEDIMENTS

Many natural sediments have been found to have a size-frequency distribution which corresponds to normal probability. For this reason the sand mixtures produced by the methods suggested in this paper may be considered to be the same as natural ones. Moreover, Doeglas (1946) has demonstrated that apparantly abnormal distributions are nothing but remnants of normal ones from which the other parts have been removed by sedimentary processes. In the event that a skew distribution is desired, the present procedure may be modified in accordance with the data presented by Otto (1939).

THE PREDETERMINED PARAMETERS

There are any number of normal probability curves, but only one which has a certain mean and standard deviation (Krumbein, 1936). This is evident if it is remembered that the curve can be plotted as a straight line on probability paper, and that the mean and the standard deviation represent two points on that line. Therefore, it is necessary to consider only these two parameters if Krumbein's phi notation is to be used.

The coefficient of sorting and the me-

dian diameter also establish a unique probability curve. In fact the mean and the median are equal if the curve is normal (non-skew). Further, it may be demonstrated by use of tables for the solution of the normal probability curve, that the coefficient of sorting (C_s) is related to the phi standard deviation (σ_ϕ) as follows:

$$Q_1 - Q_3 = 1.349\sigma_\phi$$

and since $$C_s = \sqrt{\frac{Q_1}{Q_3}}$$

$$C_s = \sqrt{\frac{Q_3 + 1.349\sigma_\phi}{Q_3}}$$

$$= \sqrt{1 + \frac{1.349\sigma_\phi}{Q_3}}.$$

Krumbein (1939) has indicated the same relation by noting that the phi quartile deviation equals 0.6745 σ_ϕ. In short, either a sorting coefficient and median, or a standard deviation and mean will define a unique curve.

THE CURVE OF PROBABILITY

The equation of the "bell-shaped" probability curve has appeared in this journal at least twice (Krumbein 1938, Otto 1939), and extended discussion here would be pointless. The solution of the integral for the area under the curve is an infinite series. The solution for the area under a part of the curve, which is the exact equivalent of the weight per cent of a sample, may be obtained from tables in any text on statistics. One advantage of selecting the phi standard deviation instead of the coefficient of sorting as a parameter is that the former may be used directly in these tables whereas the latter requires some computation.

USE OF THE NOMINAL MEAN DIAMETER

A desired sediment may be synthesized by mixing together various weights of various grain sizes obtained by sieving a natural sediment. However, the pertinent grain size is the nominal phi mean diameter ($M_{\phi n}$) rather than the diameter of the sieve openings. The nominal phi mean diameter is equal to the sieve opening plus one half of the sieving interval. Its use results in a plot on probability paper which is symmetrical about the sieve size which contains the nominal mean of the whole sand mixture. By this device, the weight percentage of any grain size can be located directly below the size on the graph presented in this paper.

The nominal phi mean diameter is only an arithmetic approximation of the true phi mean diameter of the grains which are separated by a given sieve. In a natural sand, the true phi mean diameter of these grains will vary according to the normal probability distribution. It may be obtained from a probability graph by taking the average of the cumulative weight percentages of the subject sieve size and that of the next larger sieve. The diameter which corresponds to this average weight percentage for the given sorting is the true mean diameter. It may be calculated by the use of tables for the normal probability curve if the graph cannot be read to the desired accuracy.

The difference between some true and nominal phi mean diameters is shown in Figure 1. The sieve interval is one phi unit. It can be seen that for a given sieve size the difference increases as the standard deviation (or sorting coefficient) becomes smaller; and that for a given sorting the difference increases as the dispersion of the sieve size from the mean sieve size increases.

This difference introduces little error if the sand mixture to be synthesized has a size-frequency distribution similar to the natural sand from which the sand fractions are obtained. The error is negligible, regardless of the mean size, if both the sands have the same M_ϕ, if $\sigma_\phi \geqq 2$, and if the sieve interval is no greater than one phi unit. This is shown in Figure 1 in that the true and nominal

mean diameters are almost equal within the range of standard deviations specified. The difference becomes important as the sorting improves. For example, it is impossible to produce a sand mixture with a normal probability distribution and a phi standard deviation of 0.5 by mixing size fractions derived from a natural sand with a $\sigma_\phi = 4.0$, and with the same mean. The relation can be seen in Figure 1 by considering the sand fraction which has a nominal M_ϕ larger by one phi unit than the M_ϕ of the whole sand. The fraction from the natural sand has a true mean diameter of about $M_\phi + \phi$, but the desired sand fraction has a true M_ϕ of about $M_\phi + 0.7\phi$. Thus various weight fractions of the natural sand may be mixed together according to the proportions for the desired sand, but the mixture so synthesized will not have a normal probability distribution.

There is a further complication because the mean of a natural sediment seldom coincides with the nominal mean diameter of the mean sieve. However, if the difference is no more than $\phi/10$, this error may be neglected. For fractions other than the mean one, this complication usually may be neglected regardless of the mean of the sand, provided the sieve interval is no greater than one phi unit. A comparison of lines (b) and (3) in Figure 1 shows this. The former line has a true M_ϕ which coincides with the nominal M_ϕ in the mean size fraction, however, for the latter one the true and the nominal M_ϕ differ by $\phi/2$. These two lines, therefore, represent the maximum possible range of differences be-

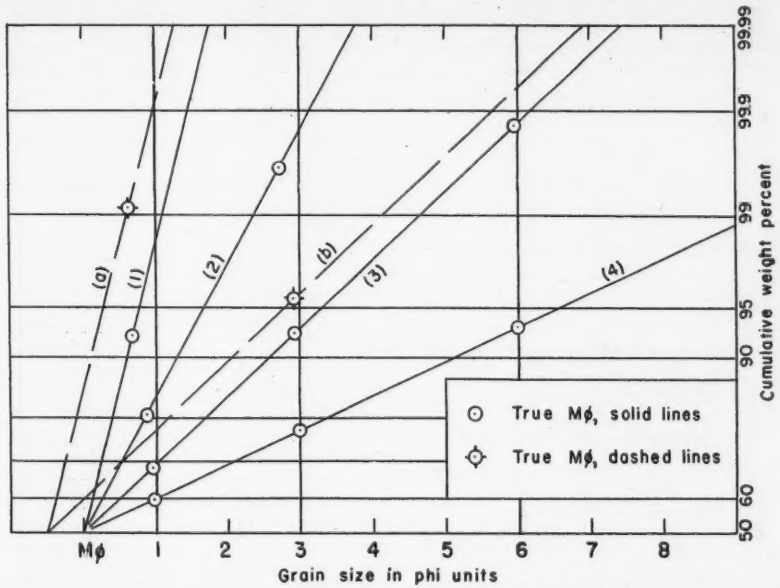

FIG. 1.—Variation between the true and nominal phi mean diameters. Size-frequency distributions are normal. Solid lines (1–4) indicate that the true and nominal phi mean diameters are equal in the mean size fraction; dashed lines (*a–b*) that the two differ by $\phi/2$ in the mean size fraction. Vertical lines through the grain diameter numbers show the nominal M_ϕ for that size fraction. The horizontal distance between the true and the nominal M_ϕ in any size fraction represents the variation in phi units.

tween true and nominal M_ϕ for a sieve interval of one phi unit. Nevertheless the true mean diameters of the lines are very similar for each size fraction except the mean one; and thus the differences between true and nominal mean diameters in each fraction are about the same for the two lines. Lines (a) and (1) show that this relation is true of any sorting.

TABLE 1.—*Proportional parts of the various size fractions of sedimentary mixtures with a normal size-frequency distribution and various standard deviations. The numbers may be considered as weight percentages although the whole of the distribution may not fall within 14ϕ and thus the total may be less than one hundred percent.*

Dispersion from M_ϕ in phi units		σ_ϕ			
		0.5	1.0	2.0	4.0
	7			0.000492	2.170
	6			0.00242	3.245
	5		0.000338	0.92	4.575
	4		0.02296	2.79	6.050
	3		0.5967	6.55	7.515
	2	0.135	6.06	12.10	8.790
	1	15.735	24.17	17.47	9.640
	M_ϕ	68.26	38.30	19.74	9.950
	1	15.735	24.17	17.47	9.640
	2	0.135	6.06	12.10	8.790
	3		0.5967	6.55	7.515
	4		0.02296	2.79	6.050
	5		0.000338	0.92	4.575
	6			0.00242	3.245
	7			0.000492	2.170

MATHEMATICAL SOLUTION OF THE PROBLEM

The solution for the area under a portion of a normal probability curve may be obtained from tables in most texts on statistics, as has been noted before. This area is presented as a fractional part of the whole area, and is exactly equal to the weight percentage of a sedimentary size fraction in relation to the whole sample. The area under the curve is usually divided into fractions by measuring distances along the X axis in terms of the standard deviation. Table 1 summarizes the weight percentage of each grain size necessary to synthesize a sand mixture with a probability size-frequency distribution and certain definite standard deviations. The sieve interval is equal to one phi unit, but the table applies to any mean size.

GRAPHICAL SOLUTION

A graphical solution to the present problem is easily constructed by the use of probability paper (Krumbein and Monk, 1942). The size-frequency distribution is a straight line, and the mean and standard deviation are two points

on that line. Figure 1 represents the plots of normal sands with a variety of sortings. Sorting is expressed in terms of phi standard deviation, and grain size in intervals of phi units. Only the upper half of the graph is evident, but the lower half may be read by inverting the page and subtracting the ordinate from 100%. Procedure for determining the weight per cent of any grain size can be shown by using $\sigma_\phi = 1.0$ as an example. The size fraction containing M_ϕ includes all diameter sizes within $\phi/2$ on either side of M_ϕ (if $M_\phi = M_{\phi n}$, and the sieving interval is one phi unit). The weight percentage which corresponds to a nominal mean diameter of $M_\phi + \phi/2$ is about 69 for this sorting.

69 (weight percentage of $M_\phi + \phi/2$)
-50% (weight percentage of M_ϕ)
19%

This percentage must be doubled to take account of the lower half of the curve. The product, 38%, agrees well with the calculated value of 38.3% shown in Table 1. The size fraction which has a $M_{\phi n}$ one phi unit larger than the $M_{\phi n}$ of the mean fraction includes all the grain sizes within the limits of $M_{\phi n} + \phi/2$ to $M_{\phi n} + 3\phi/2$. Applying the method indicated above;

93% (weight percentage of $M_{\phi n} + 3\phi/2$)
-69% (weight percentage of $M_{\phi n} + \phi/2$)
24.5%

Therefore, 24.5 is the approximate weight percentage of a size fraction which has a nominal mean diameter larger by one phi unit than that of the mean size fraction. Since the distribution is symmetrical about the mean, this is also the weight percentage of the size fraction which has a $M_{\phi n}$ smaller by one phi unit than that of the mean size fraction.

An analogous graph can be prepared relating the nominal median diameter in millimeters and the coefficient of sorting to the weight percentage. It is necessary to use logarithmic probability paper instead of the arithmetic type. The straight line corresponding to a desired sand mixture can be drawn by connecting the median to one of the quartiles. The former is given, but the quartile must be calculated from the given coefficient of sorting. This can be done as follows:

$$C_s = \sqrt{\frac{Q_1}{Q_3}}$$

therefore

$$Q_1 = C_s^2 Q_3.$$

The two quartiles are symmetrically disposed around the median so:

$$Md = C_s Q_3$$

$$Q_1 = C_s Md.$$

The weight percentages of each size fraction necessary to synthesize a desired mixture can be determined by use of this type of graph in exactly the same way as was done with Figure 1.

LABORATORY PROCEDURE

The required size fractions may be obtained by sieving a natural sand if only a small quantity of a synthetic sand mixture is desired. Yet if several cubic feet of sand are required, it will usually prove easier to modify a natural sand rather than to sieve all of it. This may be done by sieving part of the sand in order to determine its size-frequency distribution, and also to obtain some quantity of each of the desired grain size fractions. The weight percentage of each size fraction of the natural sand may then be modified by the addition of more or less of the sieved sand until the desired weight percentage of that fraction is obtained. As an example, 100 grams of a sand with a phi standard deviation of 0.5 can be changed to 100 grams of one with the same mean but with a $\sigma_\phi = 1.0$ as shown in Table 2 (data of first two columns from Table 1). Since only 38.30 grams of grain size M_ϕ are required, only 38.30/68.26 ($=0.561$) of the original 100

grams can be used. These 56.1 grams contain the percentage weights for each size fraction which appear in column (3) of Table 2. The weight of each size fraction required to be added to the 56.1 grams of the natural sand in order to synthesize the desired one can be obtained by subtracting the figures in column (3) from those in column (2). The results appear in column (4).

This method of modification may be particularly useful if it is desired to produce a large number of sands of various sortings; each one can be modified from the one with the next better sorting. Some limitations on the method have been mentioned under the discussion of the true and nominal mean diameters. Thus the example of modification selected would be undesirable because the sieve interval of one phi unit is too large to make the true and nominal mean diameters reasonably equal for all size fractions. The limits of error in any proposed experiments would govern the acceptability of any modification.

The actual physical mixing of the sand must be done so as to insure homogeneity. This may be accomplished by the methods commonly used to obtain a composite sample.

USES FOR SYNTHETIC SAND MIXTURES

Synthetic sand mixtures should find their greatest use in the experimental laboratory. They make it possible to vary the mean and the standard deviation in a systematic manner, and thus to evaluate the effect of these variables upon the process under consideration. Krumbein and Monk (1942) used such mixtures in measuring the influence of sedimentary parameters upon permeability. The author used them in an attempt to determine the relation between sorting and competent velocity.

However, other uses appear possible. Emery (1948) has shown that rapid size analyses of a great many specimens may be achieved by visual comparison with natural sediments which have a known size-frequency distribution. Synthetic sands should provide an easy means of obtaining a known suite in which each specimen varies from the next in accordance with a predetermined limit of accuracy rather than the chance variation in a natural suite.

TABLE 2.—*The numbers are weights in grams. Columns (1) and (2) represent normal size-frequency distributions for the standard deviations indicated. See text for significance and use of table.*

Dispersion from M_ϕ in phi units		(1) $\sigma_\phi = 0.5$	(2) $\sigma_\phi = 1.0$	(3) $0.561 \times (1)$	(4) (2) − (3)
	4		0.02		0.02
	3		0.60		0.60
	2	0.135	6.06	0.076	5.98
	1	15.735	24.17	8.83	15.34
	M_ϕ	68.26	38.30	38.30	0.0
	1	15.735	24.17	8.83	15.34
	2	0.135	6.06	0.076	5.98
	3		0.60		0.60
	4		0.02		0.02

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JOURNAL OF SEDIMENTARY PETROLOGY, VOL. 19, NO. 2, PP. 78–81
FIG. 1, AUGUST, 1949

MISSOURI RIVER SEDIMENTS IN RIVER WATER, OCEAN WATER, AND SODIUM OXALATE SOLUTION

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ABSTRACT

The mud suspended in Missouri River flood water was mechanically analyzed by measuring its sedimentation in river water, N/100 sodium oxalate solution, and in saline ocean water. Radically different results were obtained from the three different liquid suspensions. The "correct" mechanical analysis of an argillaceous rock may be debatable. Analytical data may hold value in interpreting the history of the rocks.

INTRODUCTION

The particle size data obtained from mechanically analyzed, disaggregated clays and shales have been interpreted with varying degrees of significance. Optimists have suggested that the particle size analysis of a shale represents the sizes of the clay and silt particles which settled down to form the rocks. Students on the other extreme refuse to credit the analyses with being much more than a measure of the efficiency or extent of the disintegration and dispersion operations, and doubt if an analysis has any worth equivalent to the time needed to perform it. The prime question involved is that of the fidelity of the mechanical analysis.

A non-conventional approach to the question presented itself to the writers while working on some muddy flood water of the Missouri River in 1947. The naturally dispersed sediment was used as a constant factor and the environment during sedimentation was varied from the conventional procedure to simulated natural conditions. It appears that a conventional mechanical analysis of an argillaceous rock may have interpretive value, but it probably is not a very reliable indicator of the sizes of the original particles in the sediment.

EXPERIMENTAL WORK

Five gallons of muddy flood water were taken well out in the channel, at the surface, of the Missouri River at Jefferson City, Missouri, on April 17, 1947, by Mr. Hal G. Stephens. Although a still higher record flood peak was reached later in the year, the sample collected was representative of the load of the Missouri River during times when it is contributing generously to modern sediments. The pH of the fresh muddy water was 7.7, whereas after filtering, the clear portion tested 7.98.

Three 500 ml. portions of the muddy river water were taken for sedimentational analysis. One was analyzed "as is," i.e., without further treatment or addition except that after siphoning off the clay size fraction in suspension, filtered river water was added to return the volume of liquid in the sedimentation cylinder to original 500 ml quantity. River water was therefore the only suspending liquid.

The second portion was shaken with sodium oxalate to make it $N/100$ in sodium oxalate. The same chemical environment was continued through the analysis.

The third portion was evaporated slowly to a volume of 200 ml. At the same time 500 ml of natural ocean water taken south of San Francisco were evaporated to 300 ml. The two portions, totaling 500 ml were quickly shaken together and sedimentation allowed to

take place, simulating a Missouri River being run into the ocean without appreciable dilution of the "infinitely large" ocean of saline water. In nature there is mixing of sweet and saline water over a highly variable range, so variable that no attempt was made to simulate it in the laboratory experiment.

Prior to the sedimentational analysis, each sample was sieved wet through standard Tyler screens having openings of the Wentworth size scale, and the sand sizes, greater than .061 mm in diameter, were scalped off. After sieving, the suspended fraction was run into a sedimentation cylinder and the clay fraction, less than 1/256 mm in diameter, was drawn off in the conventional manner. Settling velocities in the sweet and heavy saline waters were computed from Stokes' law, rather than from Wadell's formula (Krumbein & Pettijohn), because comparison might be made directly with older analyses also based on Stokes' law. The sand, silt, and clay fractions found in each 500 ml sample are shown in the table below in terms of percentage of total sediment.

	1 River Water	2 Na Oxalate Sol.	3 Ocean Water
On 0.495 mm	Tr.	0.69	0.14
On 0.246 mm	0.22	0.42	0.14
On 0.124 mm	0.19	0.27	0.20
On 0.061 mm	1.47	1.85	1.18
Total Sand	1.9	3.2	1.7
1/16–1/256 mm, silt	35.3	25.3	82.0
Less than 1/256 mm, clay	62.6	71.4	16.4
	99.8	99.9	100.1

DISCUSSION

The variation in percentages of silt and clay between the different analyses is striking, and is considerably greater than could be attributed to experimental error. Presumably the analysis in No. 1 is most nearly correct for that represents the sediment in its original and natural environment. A deposit from it on the Missouri River flood plain (and there was considerable of it—see photo of mud cracks Fig. 1) supposedly would have that size distribution until or unless diagenetic processes altered the particle sizes.

However, suppose an ambitious sedimentary petrographer were to collect the newly deposited flood plain sediment and analyze it by the conventional procedure, using Na oxalate as a deflocculant, would not the analysis have been reported as that under column No. 2? If so, the quantity of clay reported would be incorrect by about 14% too high, and the silt by about 28% too low. If the sample had been disintegrated by grinding which was a little too drastic, the errors would have been greater in the same directions. Mechanical analyses of argillaceous rocks obviously may depart widely from expressing the original sizes of the constituent particles.

Still more striking is the deviation of the data in column No. 3 from those in the other two columns. About 46% of the sample (nearly half of it), or expressed another way, two thirds of the clay was flocculated by the sea water to become silt-sized particles. Just what is the "correct" mechanical analysis of a river sediment which is being flocculated by ocean water may be open to debate. Perhaps a correct analysis would have little value anyhow.

The recognition that much marine silt may originate from flocculated clay-sized particles and clay colloids may be of greater significance. Possibly marine shale and clays are coarser-grained and richer in silt sizes than are analogous fresh water deposits, unless diagenetic

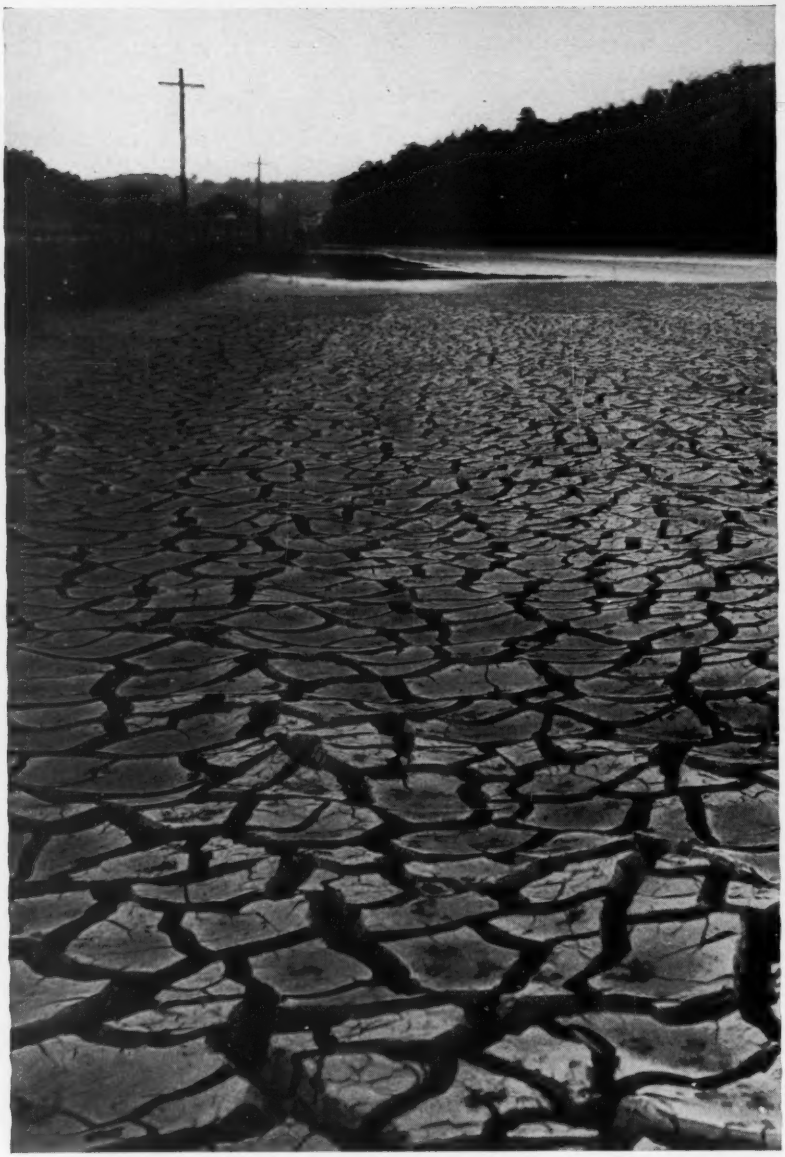

FIG. 1.—Deeply cracked, river mud deposit, at least six inches thick, left by Missouri River flood, 1947. (Near Jefferson City, Mo.)

or still later recrystallization alters the primary texture of either. Research on this point by starting with a study of the more recent sediments and working toward the older ones may be productive.

Laminated structure of shale is thought to arise from compacted platy to linear silt particles which originated from the flocculation of clay (Keller, 1946). Clay partings in limestone have been ascribed to wide-spread flocculation of clay colloids (Keller, 1936).

The progress of some diagenetic changes may be traced by analyzing mechanically a series of progressively older and more compacted field samples, and by laboratory simulation of nature necessary to duplicate the changes.

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FIGS. 1–5, AUGUST, 1949

RIPPLE MARKS AS AN AID IN DETERMINING DEPOSITIONAL ENVIRONMENT AND ROCK SEQUENCE

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ABSTRACT

Ripple marks, both wave-formed and current-formed, are of value in determining the environments under which sediments are deposited and also the sequence of their deposition. Both kinds of ripples form on stream bottoms as well as on the bottoms of lakes and other bodies of water.

A number of important differences are given that help to distinguish between wave-formed and current-formed ripples. Also, attention is called to several types of ripple marks that can be used in determining sequence of rock layers.

INTRODUCTION

Ripple marks, caused by both water currents and by wave action, are formed beneath the waters of lakes and seas as well as on the bottoms of streams. The two classes of ripples may occur separately or may be superimposed on each other, depending on the conditions prevailing at the time of their formation. Since ripple marks are often a help in determining the conditions under which sedimentary rocks are laid down as well as furnishing a key to their present sequence, it is important to be able to distinguish between the different kinds of ripples and to recognize the many forms in which they occur. At first thought, this would seem to be a relatively simple matter. It is not difficult if the exposed areas are large, and if the ripples have been formed under conditions of uniform depth of water and a constant direction and velocity of water current or wave travel.

But, unfortunately, ripple marks are often formed under complicated conditions of changing wave size and variation in depth of water or by currents that are in the process of change in both direction and velocity. Thus, detailed and careful studies are often necessary to determine whether an exposure of ripple marked sediments was laid down in a river, along the shores of a lake, or beneath the deeper waters of the sea.

DIFFERENCES BETWEEN WAVE-FORMED AND CURRENT-FORMED RIPPLE MARKS

In distinguishing between wave-formed and current-formed ripple marks, it is helpful to keep in mind that the results of rhythmic wave movement are usually much more uniform over a given area than the movements caused by flowing water. Also that a mass of sediment acted on by running water tends to be streamlined and more drawn out in the direction of the moving current than is a mass of sediment acted on by the waves.

Current-formed ripples, except when produced at certain critical velocities, are always asymmetrical. It has been stated repeatedly in our textbooks that wave-formed ripple marks, since they are the result of oscillatory movements on the bottom, are always symmetrical. Unfortunately, this is an error (Evans, 1941). Within the breaker zone, the deformed waves set up oscillatory movements that are stronger in one direction than the other. Consequently, in the shallow water along the shores, symmetrical and asymmetrical ripple marks are about equally numerous.

Thus, in studying fossil ripple marks,

it is necessary to look much farther than the question of symmetry in attempting to distinguish one class from another. Also, in making such studies, it is important to examine the exposures over as large an area as possible, because currents strong enough to cause ripple marks are often present in lakes and oceans, and waves large enough to produce oscillation ripples are common on the larger rivers. Hence, the two kinds of ripple marks may be found in close proximity or even superimposed, one on the other.

On account of the difficulties and complications of the problem, it is important that the student, before attempting to give a meaning to the ripple markings in consolidated sediments, make a series of careful, systematic observations of ripple marks as they are being formed in the streams and lakes of the present time. In doing so, the following contrasting differences should be noted.

1. *Differences in the shape of the longitudinal axes.* In wave-formed ripples, the longitudinal axes tend to be nearly straight and parallel for long distances or, if curved, the bends are gradual and uniform. A series of such ripples appear nearly parallel and show a quite uniform pattern over wide areas. This is because of the uniformity of the wave movements that produce them. In current-formed ripples, the longitudinal axes are crooked and irregular with many sharp angular turns, and the parallelism and regularity of pattern is much less noticeable than in the wave-formed ripples. This is because, along the cross-section of a stream, the velocity of the current is constantly changing. In some places the sand grains are hardly moved at all, in others they are pushed rapidly forward by the swiftly moving water.

Sometimes these differences are alone sufficient to distinguish between the two types of ripple marks. But in shallow water of varying depth, both wave size and wave direction may change rapidly, causing the axes of the ripples to be crooked and irregular. Such wave-formed ripples may be easily mistaken for the current type, especially if the area examined covers only a few square feet. And the situation is especially deceiving if there has been a change in wave direction that has partly destroyed one set of ripples and begun the formation of another.

Another misleading appearance is sometimes brought about by the slow disintegration of wave-formed ripples. Heavy storms produce ripple marking on the bottom in relatively deep water and such markings are fairly persistent. But ripple marks formed by one set of waves will be quickly destroyed if subjected to the action of larger waves, especially if there is a change in wave direction. But ripple marks formed at considerable depths by large waves may be subjected to the action of smaller waves for a long time with no appreciable change of form. However, after a considerable time they do gradually break up. In the process, their longitudinal axes become crooked, their spacing becomes irregular, and they finally come somewhat to resemble ripple marks formed in shallow streams by slowly moving water. In some of the clear, sandy-bottomed lakes of the northern states, I have often seen, in the summer months, such ripples disintegrating in depths of 10 to 12 feet of water. They had been formed during the heavy storms of winter and early spring but were gradually losing form through the imperceptible movements set up on the bottom by the smaller waves.

2. *Differences in uniformity in height of the crests of the ripples above their troughs.* Because of the regularity of wave action, wave-formed ripples tend to be uniform in height throughout their lengths. But current-formed ripples, because of the irregularities of stream flow, change rapidly in height along their crests. Unfortunately, this rapid change in height is also present in wave-formed ripples in deep water that are in the process of disintegrating. So, in distinguishing between the two types, it is necessary to consider other characteristics.

3. *Differences in shape of cross sec-*

tion. Except under very special and limited conditions, current-formed ripple marks are asymmetrical. But, so are many wave-formed ripple marks formed within the breaker zone, and it is sometimes impossible to distinguish, from shape of cross section alone, asymmetrical wave-formed ripples from those produced by slowly moving currents. The crests and troughs of both are alike in being considerably rounded (fig. 1). But, with increasing velocity of the water, current-formed ripple marks become sharper in both their crests and troughs. This angularity increases with increased velocity of the water until, by the time the ripples are traveling visibly downstream, their crests and troughs are sharply angular. However, the angle in the trough is sometimes slightly rounded by the eddy action that takes place on the lee side of the ridges. A distinguishing mark that is sometimes present in the wave-formed ripples is a small sharp ridge somewhere on its sides or at the crest (fig. 2). This is caused by a sudden surge of the water as the wave passes. These ridges are usually somewhere near the crest of the ripple and more frequently on the side of the short slope than on the long one.

4. *Differences in the shape of the wave-formed and current-formed ripple marks in the vicinity of obstructions*. Where flat rocks are present on the bottom, waves will sometimes continue a system of ripple marks across them with the crests clearly defined but with little or no sediment present in the troughs. These defective ripples are the result of the absence of sufficient sediment to complete the ripple structure. The sediment present is swept back and forth into the position where the crest of the complete ripple would be if the rock were absent. But if the ripples are current-formed, such flat-topped rocks are swept clean of sediment and the surrounding ripples end abruptly at its edges. However, if the rock, instead of lying with its upper side flat and even with the bottom, rises some distance above it, the wave-formed ripples retain their full size until they end abruptly against its sides. But if the ripples are current-formed, they will end some distance from the sides of the rock at the edge of a depression dug around it by the increased current velocity at the sides of the rock. Frequently the ends of the ripples near the rock are somewhat more irregular than farther away and their ends are turned down stream.

5. *Differences in lunar shaped ripples and in the pointing of the ends of their segments*. In the zone of waves of translation close to shore, the wave-formed ripples are often broken up into segments. Because the currents of the swash are stronger than those of the backwash, the ends of these segments all turn in the same direction toward shore. Being wave-formed, they have crests and troughs of nearly uniform elevation and width. Often the segment is so short that the form of the marking is lunar shaped. Sometimes the top of an offshore ridge becomes covered with ripple marks formed by small waves. Then an increase in wind velocity produces larger waves that break on the ridge and cause waves of translation to pass across it. These may cause a breaking up of the wave-formed ripples into segments, whose ends all point in the direction of movement of the current. Thus a pattern of lunar segments is sometimes found covering a considerable area. Somewhat similar horseshoe shaped ripples form on stream beds or where the current sweeps the bottom of a depression in a water body. But they are more irregular in their distribution and their ends point in either direction, against or with the current. If they point downstream, the steeper slope of the ridge is within the bend of the horseshoe. If they point upstream, the steeper slope is on the outside of the horseshoe bend.

RIPPLE MARKS OF UNIQUE FORM

Thus far we have emphasized outstanding differences between wave-formed and current-formed ripple marks. However, there are a few ripple forms so distinctive and unique that, when found

well preserved in the rocks, it is impossible to mistake the class to which they belong.

In rippled areas produced by waves, smaller symmetrical secondary ridges are sometimes present within the troughs. They are caused by a change in wave length following the formation of the original set of ripples. This ripple pattern is unlike anything in the current-formed class (fig. 3).

Also in the wave-formed class, there are two minor forms that should be mentioned, though they are rarely found preserved in sediments. One is a ripple with rounded crests and angular troughs (fig. 4). It is found in narrow belts near shore and results from the action of the slowly dying waves following a storm. Its form is exactly the reverse of the very common form of symmetrical wave-formed ripple having rounded troughs and angular crests. Another minor form is a double ripple mark that occurs in narrow belts not far from shore and parallel with it (fig. 5). It is probably caused by interfering nodes as the waves change size in approaching the shore. There is nothing else like it in either wave-formed or current-formed ripples.

A distinctive marking that is present in many stream-formed ripples is a small ridge trending in the direction of the flow of the current. Because of the great irregularity in current velocity, spoon-shaped depressions, having their long axes in the direction of stream flow, frequently form on stream bottoms. Probably they can be considered as overgrown "linguoid" ripples (Bucher, 1919). Usually they are only a few inches long but may be very much larger, sometimes reaching a length of several feet and a depth of a foot or more. Occasionally two or more of these form side by side or even overlap. This results in the formation along their edges of small streamlined ridges lying parallel with the direction of stream flow.

I have found such markings very common on the rippled bottoms of the clear streams flowing over glacial deposits as well as in the current rippled sands in the the beds of the Canadian and other rivers of Oklahoma. They are also preserved in

FIG. 1

FIG. 2

FIG. 3

FIG. 4

FIG. 5

many places in the stream-rippled sandstones so common in the Pennsylvanian and Permian rocks of Oklahoma. These ridges furnish a positive identification for current-formed ripples and are also clear indicators of the direction of stream flow. For they are always found on the upstream slope of the ripple, beginning in the depression at the foot of the ripple above and ending at the crest of the ripple on which they lie.

USE OF RIPPLE MARKS IN DETERMINING SEQUENCE

In the above, I have indicated some of the things to be considered in using fossil ripple marks in determining the environment in which the sediments are deposited. In deciding whether or not rocks have been overturned, a little consideration is necessary to determine which forms can or cannot be used. Thus, it is at once evident that, to be of value in determining sequence, a ripple mark and its mould must be unlike. Thus, symmetrical wave-formed ripple marks having secondary ripples in their troughs are completely reliable guides since there are no forms of ripples the reverse of them. Also, there are no forms that are the reverse of the double ripple marks (fig. 5), or of the wave-formed ripple marks having a slight ridge somewhere on the crest (fig. 2). Again, symmetrical wave-formed ripples with sharp crests and rounded troughs (fig. 4) are good indicators but are not, of themselves, absolute proof of sequence since, as we have seen, there are sometimes narrow belts of ripples formed in the breaker zone near shore just their reverse, with angular troughs and rounded crests. Among the stream-formed ripples, the streamlined ridges found on the upstream slope of some of the ripples can be depended on in determining sequence since there are no ripple forms the reverse of them.

However, the above is only a brief guide. Ripple marks are of so many forms and there are so many possible combinations of these forms that anyone proposing to use ripple marks as a help in studying either depositional environments or sequence of the formations should first spend several days or weeks in the field, observing modern stream-formed and wave-formed ripples. It is a thing in which the eye as well as the intellect needs training. It is as difficult to become a good ripple reader without practice as it would be to become an expert hunter and trailer by merely reading a book on the habits of animals.

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THE ORIGIN OF THE VERDEN SANDSTONE OF OKLAHOMA[1]

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ABSTRACT

The Verden sandstone is an elongate formation that outcrops in a series of buttes and ridges from about 10 miles east of Marlow, Oklahoma, northwest for about 70 miles. It has a thickness of 10 feet and an average width of 1,000 feet. Its beds, which vary from a few inches to two feet in thickness, are extensively and persistently cross-bedded toward the northwest. These beds are separated in many places by thin, horizontally laminated sandy shale, that contains both wave-formed and giant ripple marks having their longitudinal axes approximately parallel with the trend of the formation. The sediments become progressively finer to the northwest and marine fossils, that show only a small amount of wear from transportation, are common.

Of the several theories that have been suggested to explain the Verden, the one advocated by Bass (1939), that it was laid down as a series of spits and bars, has been most widely accepted. However, in the light of our present knowledge of the nature and internal structure of elongated physiographic formations, it seems that the peculiarities of the Verden sandstone are better explained as caused by salinity or tidal currents flowing through a strait or pass.

DESCRIPTION OF THE FORMATION

The Verden sandstone is a long, narrow body of rock extending from 10 miles east of Marlow, Oklahoma northwest nearly to Calumet, a distance of some 70 miles. Throughout a considerable part of its length, it stands out prominently in a series of buttes and ridges, and has been mentioned by a number of writers on Oklahoma geology.

The most important and extensive of these studies was made by Bass (1939). In the abstract of his article, he describes the formation as follows: "The Verden sandstone body is only 10 feet thick and less than 1,000 feet wide in most places. It lies in the redbeds portion of the Permian series. The sandstone occupies everywhere the same stratigraphic position within a range of a few feet, in the redbeds, and lies within a sequence of regularly bedded thinly laminated silty sandstones, silt, and gypsum of the Marlow formation. It is characterized by relatively thick beds of medium- to coarse-grained sandstone which are steeply cross-bedded with northwest dips, and are interbedded with beds of fine-grained horizontally laminated sandy shale. The sand grains have a wide range in size but are sorted into layers in which grains of a single size predominate. The fine grained beds contain wave ripple marks and giant ripple marks that trend nearly parallel with the sandstone belt. The sandstone is a firm hard rock composed of well rounded and subangular chert grains and of calcium carbonate which makes up 50 per cent of the rock. Marine fossils are common."

The great length of the Verden sandstone compared with its width places it in that class of deposits that are variously designated as "trends," "channel sands," or "shoestring sands." But, in its internal structure, the Verden sandstone differs from all other shoestring sands that have been described. Throughout 50 miles of its length, from where it outcrops about 5 miles northeast of Rush Springs to beyond the South Canadian River, it is persistently cross-bedded to the northwest. And nearly everywhere, these beds are separated by thin, horizontally laminated sandy shale beds that range from paper-thin up to

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several inches in thickness. Bass (1939 p. 570) describes the cross-bedding as follows: "The cross-bedded layers dip at angles ranging from less than 1° to 24°, but commonly range from 10° to 20°. Although the predominant dip of the cross-bedding is northwestward in all exposures observed, the directions of dips in the several beds that make up the sandstone at each single exposure range through a rather broad arc. The writer observed local examples of cross-bedding that dip into each of the four quadrants of the compass. For instance, in the SW ¼ of Sec. 13, T. 7 N., R. 9 W., a mile southwest of Verden, the predominant dip of the beds is northwestward, but in an area 20 feet in diameter some beds dip northeast, others northwest, others southwest, and still others southeast. Moreover, throughout a distance of 200 feet on the northeast side of the butte in Section 13, the cross-bedded layers dip predominantly southwestward, but elsewhere near by they dip northwestward."

It should be noted that the characteristic cross-bedding of the Verden sandstone is absent at both ends of the formation. The two southernmost of the outcrops, about 10 miles east of Marlow, are isolated by a distance of 12 miles from the nearest Verden to the northwest. The writer saw only the larger of the two, the one in the middle of Sec. 13, T. 2 N., R. 6 W. This outcrop is like the Verden lithologically but only a few of its layers are cross-bedded, not enough to have special significence, though what cross-bedding there is, dips northwest. Also, at the north end of the Verden, beyond the center of T. 12 N., R. 9 W., there is very little cross-bedding.

On Bass' map (1939, p. 561), several outcrops are shown in T. 9 N., R. 9 W. which, except for those in Section 5, he did not see personally. They are mostly Rush Springs sandstone instead of Verden. But the outcrops on the west side of Section 19 are Verden as are the ones in Section 5. Also in the southwest corner of Section 13 is an exposure, mentioned by Stephenson (1925), that appears to be Verden. This is 5 miles east of the other outcrops in the township. Again, at about the middle of the north side of Sec. 12, T. 10 N., R. 9 W. is another that was called to the writer's attention by Mr. Leon Davis of the Oklahoma Geological Survey. Like the outcrops east of Marlow and at the north end of the formation, these two are only a little cross-bedded. Possibly these widely separated outcrops mark a division in the channel in which the Verden was deposited.

Besides the persistent cross-bedding and the presence of horizontal layers of sandy shale between the beds, the Verden sandstone has three other outstanding characteristics that must be explained by any proposed hypothesis of origin. They are the increasing fineness of the sands toward the northwest, the character of the ripple marks, and the condition and position of the fossils in the beds.

FIVE POSSIBILITIES OF VERDEN ORIGIN

There are 5 well known physiographic structures that have the general elongated shape of shoestring sands. They are (1) stream channels, (2) offshore bars, (3) the "balls" or ridges of the offshore deposits known as low and ball, (4) a series of spits and bars nearly in alignment, and (5) deposits laid down by currents in the passes between islands or between an island and the mainland. Judging from its shape alone, the Verden sandstone could be any one of these.

PREVIOUS SUGGESTION AS TO ORIGIN

Reeves (1921) thought the Verden might be the result of a strong tidal bore moving up the course of a southeastward flowing Permian stream. Reed and Meland (1924) explained it as caused by a single great flood rushing northwest through one of the distributary channels of the delta of a desert river. But the evidence is strongly against the Verden being an old stream bed.

Marine fossils that show little wear from transportation occur in 18 localities (Bass 1939, p. 572), and the cross-bedded sand layers with the horizontally lami-

nated shaly beds between are too uniform to have been formed in a river. A stream swings from side to side, at the same time its curves migrate downstream. This causes secondary structural features, such as ripple marks and cross laminations to vary in their directions through a wide arc, and the turbulence and unevenness of the flow makes the cross-bedding local and irregular in occurrence and prevents the formation of thin horizontal beds over any considerable distance. There are some old stream beds now exposed in the Pennsylvanian sandstones of northeastern Oklahoma. In them, it is not unusual to find two layers of ripple-marked and cross-bedded sandstones a few inches thick; the lower one clearly showing a flow to the northeast while the lower one shows that a current ran to the southwest, yet the actual course of the stream was northwest.

A short time before making his study of the Verden sandstone, Bass (1937) published the results of his well-known work on the shoestring sands of Kansas and Oklahoma, in which he reached the conclusion that they are buried offshore bars. With respect to their internal structure he says, "A few cores showed that the thick beds of sand are locally cross-bedded at angles of 10° or less; most cores reveal horizontal beds." But, as Bass at once recognized, the Verden sandstone cannot thus be explained as an offshore bar because its prominent cross-bedding indicates that a controlling factor in its deposition was a current running almost continuously to the northwest throughout its length and this is not in accordance with the generally accepted explanation of the formation of offshore bars by the shoreward plunge of the breakers (Johnson, 1919).

Neither can the Verden have been formed as one of the ridges of the low and ball for, according to studies by Evans (1940), these are formed by sediments thrown outward from shore by the plunging breakers and so must have almost the same internal structure as offshore bars. Thus, quite naturally, Bass was led to consider the Verden sandstone as having been deposited in a series of spits and bars. He says (1939, p. 576), "The Verden sandstone was deposited as some type of a bar, or a closely related series of bars; it probably represents a spit, or a closely related chain of spits, that was extended across the mouth of a bay on the shore of a shallow sea," and he explains the cross-bedding and the presence of the silty material between the beds as resulting from the action of "longshore currents that were active periodically during storms."

CHARACTERISTICS OF THE VERDEN NOT LIKE THOSE OF SPITS AND BARS

Spits and bars are explained in our textbooks as the result of material carried by longshore currents. Some authors, following Gilbert's original illustration, liken their growth to that of a railway embankment that is constantly having material added by dumping at the outer end. But Evans (1942) has shown that the building of spits and bars is not nearly so simple a process as was formerly believed. Spits and bars, instead of being built by more or less uniformly flowing longshore currents, owe their origin to a set of much more complicated processes. An appreciable amount of the sediment in the submerged part of a spit or bar, as well as nearly all that part above the water is brought to it by the process of beach drifting and the structure increases in length only when the waves and currents together operate in such a way as to cause the beach drifted sediment to move outward toward its free end. As the waves traveling along the face of the spit reach its outer end, they refract around it, carry the sediment along with them, and so widen it. Thus the width of a spit depends on the angle at which the waves reach the shore and the amount of sediment available as compared with their transporting power. But when the wind is in such a direction as to cause beach drifting from the free end of the spit toward its base, the spit

is shortened and, at the same time, rapidly widened. Also, it is evident that, at all times, the sediment at the submerged end of the spit is subject to disturbence down to a depth equal to the effective wave base of such waves as may be present.

Considering the complexity of these processes, we would hardly expect a spit or bar to be made up of a regular series of almost horizontal beds and all of them uniformly cross laminated in one direction, as is the Verden sandstone. Nor would we expect such beds of coarse sandstone to be uniformly interbedded with layers of fine sandy silt. Instead, the bedding would be irregular, and such cross-bedding as is present would point in various directions, and whatever fine material that may have settled during calm weather would be thrown into suspension again by the breaking waves and mixed with the other sediments.

THE VERDEN A POSSIBLE CHANNEL DEPOSIT

This leaves one remaining possibility to be considered, that of deposits laid down in currents flowing through passes between islands or between islands and the mainland. Numerous examples of this, caused by the tide, are found along the British Columbia coast. According to the U. S. Current Tables, in the passes along this shore, velocities of 4 or 5 knots are common and sometimes it runs up to 10 or 12 knots.

If the movements were equal in both directions during the ebb and flow, the sediments would be swept alternately back and forth and any sand beds laid down would probably be cross-bedded in opposite directions in their alternate layers. But according to the evidence of observers, in some of the passes of the British Columbia coast, the tidal currents do not always alternate equally with the ebb and flow but may be very strong during one phase of the tide and extremely weak at the other. Thus, according to J. B. Umpleby,* former head of the Department of Geology of the University of Oklahoma, who has had considerable experience along the coasts of Washington and British Columbia, there is an almost constant current to the west through Peavine Pass, during both the ebb and flow of the tide. Also, the Pacific Coast Current Tables show the flood to be considerably weaker than the ebb in the entrance to Juan de Fuca Strait. However, Mr. T. G. Thompson,* Director of the Oceanographic Laboratories of the University of Washington, states that the figures in the current tables are accurate only for the surface waters down to a depth of 15 feet and that the currents below often differ greatly in direction and velocity from those on the surface. He ascribes this contrast between the upper and lower currents largely to the inflow of rivers along the coast combined with the effect of the prevailing winds. However, the thickness of the Verden and the character of its ripple marks indicate that it was laid down in relatively shallow water that might be under the control of the surface currents.

Another cause of flow in one direction through a pass is difference in salinity. A well known case is through the Strait of Gibralter. In the Mediterranean, the amount of water lost by evaporation is greater than that received through rainfall and the inflow of rivers combined. Consequently there is a constant inflow through the strait on the surface and an outward current on the bottom. This outward current is known to have a velocity, at times, of 4 or 5 knots. A similar condition exists in the Strait of Bab-el-Mandeb, at the entrance to the Red Sea. This is an extremely dry region, and the Red Sea has a much higher salinity than the open ocean. Thus, as at Gibralter, the lighter water from the ocean flows in on the surface and, at the same time, the heavier salt water from the Red Sea moves out on the bottom.

Also, a constant movement of sea water in one direction can be caused by differences in time and height of tide on

* Letter, 1948.

two nearby coasts. As is well known, this is the condition at the Isthmus of Panama. At the Atlantic end of the canal, the tide rises and falls less than a foot on the average with high and low water coming once a day, but at the Pacific end, the rise and fall is 12 to 16 feet and takes place twice a day. If a sealevel canal were cut through the Isthmus, there would be an almost constant flow of water from the Pacific to the Atlantic end. If the bay in which the Verden was laid down had an opening to the ocean to the south in Marlow time, it may have presented a similar situation and given rise to an almost constant current through to the practically tideless inland sea to the northwest.

Thus it is clear that straits and passes through which there is a fairly constant flow of sea water in one direction on the bottom are fairly common. And it seems reasonable to assume that if, at the inflowing end of such a strait, an abundant supply of sediment were present, conditions would be the best possible for laying down a deposit having the layers all cross-bedded in one direction as in the Verden sandstone. Bass (1939) though he finally decided on the spit and bar hypothesis, seems, at one time, to have had this same explanation in mind. For, in describing the cross-bedding and sorting of the Verden, he says, "In the coarse-grained beds it is mainly the alternation of fine- and coarse-grained bedded layers, which are accentuated by varying intensities of red, that gives the rock a cross-banded appearance. The cross-bedding is made particularly conspicuous on weathered edges of the sandstone on which the alternate fine- and coarse-grained layers stand in relief as ridges and grooves all dipping at steep angles, . . . The coarse-grained beds of the Verden sandstone are not nearly so well sorted, however, as most beach sands, but are much better sorted than most river sands. The sorting appears to be similar to that the writer has observed in beach sand at Indian River Inlet, Delaware, on the Atlantic coast. There, the local sand deposit was transported by the swift tidal currents that flow outward through the outlet."

Indian River Inlet is merely a pass between the ends of two long spits. It is very short as compared with the length of the cross-bedded part of the Verden sandstone. But examination of charts of the seacoasts shows that many of the straits and passes through which currents are known to flow are longer than the Verden. Johnstone Strait, between Vancouver Island and the mainland, is over 80 miles long and Clarence Strait, in southern Alaska, extends over 100 miles. The exact distance the currents flow through the Straits of Gibralter and Bab-el-Mandeb on the bottom are not easy to determine, but they run for at least 15 or 20 miles. It is entirely possible that today in some of these straits and passes, sediments are being laid down with a cross-bedded structure similar to that of the Verden. Not only does the hypothesis of deposition within a channel best explain the cross-bedding of the Verden sandstone but it appears to best account for its other most striking characteristics; the ripple marks, the thin-bedded horizontal layers of fine material, the increasing fineness of the sand to the northwest, and the condition of the fossils in the deposit.

RIPPLES INCICATE MARINE CHANNEL CONDITIONS

Bass found ripple marks in 37 localities. Eleven of these, he says, are wave-formed and 16 are giant ripple marks. The cause of giant ripple marks is not definitely known, but they are believed to result from currents set up by the tides; the symmetrical ones by alternating tidal currents of nearly equal strength and the asymmetrical ones by tidal currents that are much stronger in one direction than the other.

The ripple marks in the Verden, both small and large, are in the fine-grained interbedded material. None is found within the coarse, cross-bedded sandstone. This is what would be expected

in a deposit that has been intermittently swept by strong currents followed by periods of quiet water. The process of cross-bedding prevents the preservation of ripple marks within the bed itself. But any ripples formed later than the cross-bedding may be preserved during the next strong flow, because a mass of fine silt or clay, when once settled and compacted, resists disturbance by current action much more than does a coarse, unconsolidated sand bed. All the ripple marks described by Bass are wave-formed and lie with their longitudinal axes nearly parallel with the trend of the formation. He saw no current ripple marks, but in Sec. 16, T. 11 N., R. 9 W., at the southeast end of the ridge, is a well-defined exposure of current ripples. They are 3 to 4 inches wide, lie within the fine shaly material, and were made by a current flowing north-northwest with a velocity of 2 or 3 miles per hour. These ripple mark conditions are exactly what would be expected of a deposit laid down by a current in a strait or pass. The latitude of the Verden sandstone is about 35° north, well within the region of the westerlies, and its trend is almost at right angles with their prevailing direction. Had the sediments of the Verden been exposed to the open sea during deposition their longitudinal axes would point in various directions because of changing winds and wave refraction and would have formed in the coarse grained beds as well as the shaly layers.

The writer was not able to find all the giant ripples described by Bass. Some have probably been destroyed in quarrying operations. But he says some of them have their short slopes to the northeast and some to the southwest. This could have been caused by the rise and fall of the tidal waters in a strait, causing the alternating currents on the two sides to be stronger in opposite directions.

INTERMITTENT CURRENT FLOW INDICATED BY SHALY INTERBEDDING

A channel deposit most easily explains the presence of fine shaly layers lying horizontally and sharply defined between the coarser cross-bedded material. If a fine-grained deposit is laid down during quiet water in the edge of an open sea it is again thrown into suspension by the plunging breakers of the next stormy period, thus destroying both the silty layer and any ripples that may have formed in it. But in a channel, the tidal or salinity currents start gently with little or no turbulence, giving conditions most favorable for preservation and burial by the coarser material.

INCREASING FINENESS OF SEDIMENTS TO NORTHWEST SHOWS CURRENT DIRECTION

All who have written of the Verden agree that its sediments get finer to the northwest. This is exactly what it should do if deposited in a channel with currents running to the northwest and with the source of the material to the southeast. But in a series of spits and bars, the sediments may come from many places along the coast and are more likely to vary widely in degree of coarseness and kind of material.

CONFINEMENT WITHIN A CHANNEL INDICATED BY THE FOSSILS

Regarding the position and condition of the fossils, Bass (1939, pp. 572–573) says, "Fossils were observed in 18 localities distributed throughout the length of the Verden sandstone belt, but none has been found in the enclosing Marlow formation. In these localities the fossils do not occur in all beds of the Verden, but are concentrated in a few thin cross-bedded layers of local extent. The fossils consist mainly of small pelecypods and a few gastropods which George P. Girty of the Geological Survey identified as marine forms of Permian age, most of which have been collected also from the Whitehorse sandstone of Oklahoma. The valves occur separately and lie with their convex sides up, which shows that they have been transported. On the other hand, the edges of the specimens show little wear."

Again this seems to agree with its

being a channel deposit. For in a series of spits and bars the shells would have been distributed by the waves and currents through the surrounding Marlow as well as in the Verden itself and would have been fairly well worn in the process. But in a channel they would be confined to the Verden and be subjected to less breaking and grinding and an earlier, quicker burial through action of the currents than in the turbulence resulting from wave action.

Thus the evidence on all five counts favors the hypothesis of the deposition of the Verden sandstone in a strait or pass subject to tidal or salinity currents rather than its being deposited offshore in a series of spits and bars.

PHYSIOGRAPHIC ENVIRONMENT DURING VERDEN TIME

From the geological evidence available, it is difficult to reconstruct, with any certainty, the local physiography of the area at the time the Verden sandstone was deposited. The Wichita Mountains to the west and the Arbuckles to the east, were uplifted during the Pennsylvanian and were probably still relatively high lands in Verden time, and there is some evidence to indicate that a chain or group of islands may have been present between the two mountain masses during a part of Permian time.

The Verden sandstone lies entirely within the Marlow which, along with the formations below and above it, was deposited in a broad bay-like extension of the Anadarko Basin that reached southeastward between the two mountain masses. On the geological map, the area is outlined by the Duncan-Chickasha formation which appears as a great "V" with its apex lying to the southeast just beyond the southernmost outcrop of the Verden. Within this synclinal structure enclosed by the "V," is the Marlow formation and above it the younger Rush Springs and Cloud Chief. From his studies Bass concluded that the Duncan sandstone was the chief source of the sediments in the Verden. To the west and north of the Wichita Mountains a Permian sea stretched across Oklahoma into western Kansas. This sea reached south beyond the Big Bend region of Texas and is believed to have connected with the ocean across what is now the Gulf of Mexico. The buried Permian islands between the two uplifts as well as the flat-lying Permian sediments that today surround the Wichitas suggest that, at one time, the mountains lay as an island in the Permian sea.

There is a well recognized disconformity between the Duncan-Chickasha and the Marlow. It is probable that when the Verden sandstone was being deposited, the Duncan-Chickasha was exposed in the coast to the east. This is, apparently, also the supposition made by Bass. However, if the Verden was laid down in a strait or pass, there must have been also some exposure of land on the west, or at least a definite shallowing, though there is no known evidence of an unconformity in the Marlow in Verden time. Yet local disconformities in shaly deposits are extremely difficult to detect and it is possible that a low-lying land or string of islands may have been present to the west for a short time forming, with the shore to the east, a trough in which the Verden was deposited. As the Verden is only 10 feet thick, a few hundred years would have been sufficient for its formation. Such a shortlived disconformity would leave behind little evidence of its existence.

Whether the currents that carried the material for the Verden sandstone were tidal currents or salinity currents, it is impossible to say, and whether or not there was a pass through to the southeast to the open sea in Verden time, we have no way of knowing. The nature of the sediments in the Marlow and other Permian deposits of this time in Kansas and Oklahoma, suggest a dry climate. So if the inland sea in which the Marlow and the enclosed Verden were deposited happened to be fairly well enclosed, and there did exist a passage through to the

ocean to the south, conditions would have been favorable for a bottom salinity current flowing northwest, carrying material from the Duncan sandstone along with it. But there is no more direct evidence for this than there is for the existence of the embayments in the Duncan-Chickasha formation across whose mouths Bass postulated the formation of spits and bars.

Thus the evidence as to the way in which the Verden sandstone was deposited rests on the peculiar characteristics of the Verden itself; its regular cross-bedding, its interstratified shaly beds, its decrease in fineness to the northwest, the character of its ripple marks, and the kind and condition of the fossils in it. According to this evidence, it seems most logical to conclude that the Verden sandstone was laid down in a pass reaching from east of Marlow, Oklahoma to near Calumet connecting the inland sea of the Anadarko Basin with the open ocean to the south.

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REVIEW

Sequence in Layered Rocks. By Robert R. Shrock. Pp. 507; figs. 397; 34 pp. bibliography. McGraw-Hill Book Co., New York, 1948. Price $7.50.

The primary purpose of this book is to describe and illustrate the features, structures, and relations of stratified or layered rocks, which may be used to determine the original order of succession, relative age of adjacent beds, or direction of original verticality. The mode of origin of each feature is discussed to give the student the proper background to appreciate its usefulness as well as limitations. The following chapter heads indicate the scope of the book; gross relationships of stratigraphic succession (undisturbed stratigraphic and faunal succession, unconformity, igneous intrusion and comparative deformation and metamorphism), lithological- textural-mineralogical sequences, features on upper and undersurfaces of sedimentary beds, internal features of sedimentary rocks, features of igneous rocks (extrusive and intrusive), and features of metamorphic rocks.

The book is designed specifically for students, and the more than 700 references to the literature facilitates further inquiry into specific fields. The text is well illustrated with 500 excellent line drawings and more than 150 photographs. The completeness of coverage makes this book indispensable to the field geologist, regardless of whether he is working with sedimentary, igneous or metamorphic rocks. The author is to be congratulated for the tremendous amount of careful and painstaking effort necessary to bring together the vast literature on this subject into a well organized and extremely useful book.

STANLEY A. TYLER
University of Wisconsin

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NOTICES

MANAGING EDITOR APPOINTED

Professor Jack L. Hough, Department of Geology, University of Illinois, Urbana, Illinois, has been appointed by the Council of the Society as Managing Editor of the *Journal of Sedimentary Petrology*. He is to be in charge of all manuscripts, editing, preparation of copy for the printers, and advertising. Dr. W. H. Twenhofel, Editor of the Journal, is thus relieved of routine duties but will continue to be available for advice to individual authors and the council. Authors should henceforth send their manuscripts to Professor Hough.

OUT OF PRINT ISSUE REQUESTED

J. B. Alexander, Geological Survey Office, Batu Gajah, Perak, Federation of Malaya, is trying to complete his official set of the *Journal of Sedimentary Petrology* and would appreciate hearing from anyone who is willing to part with a copy of Vol. 8, No. 1, (April, 1938) at a reasonable price. This issue is out of print.

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